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The zoned contact-skarns of the Broadford area, Skye: a study of boron-fluorine metasomatism in dolomites.

(With Plates XVIII–XXIV.)

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1. Review of the progressive metamorphism in the aureole of the Beinn an Dubhaich granite.

In a preliminary note on the dolomite contact-skarns of the Broadford area, Skye (Tilley, 1948, *a*), a brief account was given of the constitution and environment of these multizoned assemblages at the contact of the Durness limestone succession with the Tertiary Beinn an Dubhaich granite.[1] No detailed survey of the numerous parageneses was then attempted, and it is the purpose of the present paper to fill this gap and to discuss the genetic problems of these skarn assemblages.

[1] The history of the discovery of these skarns and the exploratory work carried out during the 1939–45 war on some of the magnetite ore prospects has been indicated in the paper to which reference is now made.

Meantime, however, three short papers have been presented dealing with specific items of petrogenesis of the granite and its aureole: one on the earlier stages in the metamorphism of the dolomites (Tilley, 1948, *b*), on iron-wollastonite in the skarns (1948, *c*), and on the development of an alkali facies of the granite at the skarn contacts (Tilley, 1949).

The general character of the metamorphism induced in the Cambrian dolomites has been given in some detail in the Skye memoir (Harker, 1904). Three divisions of the Durness limestone succession were recognized by Harker and Clough in the mapping of this part of Skye, and these were given the local names of Beinn an Dubhaich, Strath Suardal, and Ben Suardal groups. The first two of these are commonly dolomitized while the third is normally a limestone. The Beinn an Dubhaich horizon is seen only in a highly metamorphosed state and normally underlies the other two horizons. It is not infrequently referred to as the Ring Limestone, from the occurrence of rings of forsterite (serpentine) and calcite of *Eozoon canadense* type near the granite contact.

The Strath Suardal zone is characterized by chert nodules simulating 'sponge forms', which, when not replaced, weather white, while the Ben Suardal group of limestones carries black cherts and worm casts and represents the fossiliferous horizon. Whether there are sure grounds for the separation of the Beinn an Dubhaich and Strath Suardal horizons has been debated, and for our purpose it is perhaps of no great moment since both are dolomitized. It is clear, for example, that the ring structures of the former represent the metamorphic product of a 'sponge form' which has acquired this concentric layering by metamorphic diffusion in which the 'sponge form' has disappeared or is represented only as a vestigial nucleus to the halo of rings around it.

The accompanying map (plate XXIV) reproduces with minor changes the features of the six-inch mapping of the granite and its aureole carried out by Harker. The two lower dolomite horizons have not been separately distinguished and are included together under the symbol, Strath Suardal. Upon this map have been inserted the chief localities of the skarn developments which will be discussed in detail in the sequel.

The earliest stages of metamorphism of the Strath Suardal dolomite appear on the north-east slopes of Ben Suardal and are there seen in the haloes of talc, eventually tremolite, surrounding the 'sponge forms'. These stages have been previously described (Tilley, 1948, *b*). In the inner parts of the aureole these 'sponge forms' may be completely replaced by tremolite or finally diopside, as, for example, in the strip of Strath Suardal dolomite seen at Loch Kilchrist, immediately west of the

churchyard and near the loch outlet 350 yards north of this position. In the old marble quarries enveloped by the granite magnificent specimens of ring structures surrounding 'sponge forms' now converted to diopside have long been known.

Study makes clear that the progression in metamorphism is expressed mineralogically by the successive development of talc, tremolite, forsterite, and diopside. The highest stages are represented in the structures which have their cores partially or wholly replaced by diopside, this mineral characteristically giving place to a thin forsterite zone followed by rings of calcite and forsterite in repetition.

Though this ring structure has previously been regarded as 'an original peculiarity further developed and accentuated by metamorphism', there can be little doubt that it is a diffusion phenomenon comparable with the Liesegang effect.

Already in the earliest stages of metamorphism the 'sponge forms' become encircled with haloes of talc and calcite due to diffusion, and the more striking ring structures of the inner aureole may be interpreted as a similar phenomenon developed on a more extensive scale, with the heightened amplitude of diffusion consequent upon the more elevated temperature. The eventual complete replacement of the 'sponge form' by diopside is sufficiently indicative of migration of material, and, as the form of the 'sponge' is effectively preserved, the incoming of lime and magnesia is balanced by an outward migration of silica. In the skarn zones to be described subsequently the production of analogous layered structures testifies to the efficacy of diffusion processes in these metamorphic phenomena.

Only rarely in the main body of the aureole has wollastonite been recorded, though this mineral is common at the skarn contacts. Wollastonite has so been noted in association with diopside in the dolomite of the old marble quarries, 500 yards SW. of the old Manse, Kilchrist.

The scarcity of wollastonite in the aureole of the dolomite horizons is not unexpected, as in general their bulk composition would not be such as to provide, without metasomatism, wollastonite parageneses. The dolomites have long been famous for their development of brucite-marbles in the aureole. The more detailed study of the aureole that has now been possible shows that periclase itself may be developed in force. Periclase-bearing marbles are now recorded from Kilbride and Camas Malag, particularly at the former locality. A typical example is figured in pl. XVIII, figs. 1 and 2.

Some of the brucite-marbles are characterized by a veined texture in

which a white marble carrying clear brucite is crossed by narrow veins of darker colour built of calcite and brucite, the latter turbid brown in section or spotted black. This turbidity is due to the separation of fine iron oxide from an original periclase containing ferrous oxide replacing magnesia. In the Kilbride periclase occurrences, this mineral still preserved is not infrequently light brownish in colour and gives place to a halo of brucite in which a similar turbidity and spotting is developed. The restriction of the turbid brucite to the veins of the marbles noted above suggests that the veined texture may have developed prior to or during the metamorphism and that solutions carrying iron circulated in the veins giving rise to a ferriferous dolomite subsequently to be dedolomitized. Some of these veins may carry forsterite in greater concentration than is met with in the body of the marble.

The effects of metamorphism on the Ben Suardal limestone are not so readily studied. The marbles of the skarn localities are dolomitic, and though this horizon has been originally mapped at Kilbride as contacting the granite, the innermost portion of the aureole is there dolomitic as indicated by the mineral assemblages of forsterite, diopside, and brucite, and the nature of the associated skarn minerals. Elsewhere metamorphic effects in the Ben Suardal horizon are indicated by the occurrence of grossular, idocrase, and diopside. No wollastonite has yet been recorded from this horizon. Farther north, within the mass of the Beinn na Caillich granite, an enclosure of Ben Suardal limestone has developed grossular, idocrase, and diopside. The early production of grossular garnet in this horizon in the Beinn an Dubhaich aureole is seen in the exposures of the quarry at the bend of the road 300 yards west of the west end of Loch Kilchrist.

2. The skarn zones.

The complex skarn zones now to be described occur at or close to the contact of the granite with the dolomite horizons. They are therefore essentially the metasomatic products of original dolomites. Not only do they show a layered structure against the granite, but the 'sponge' forms' which are incorporated within the metasomatic zone develop also a layered structure around the 'sponge' nucleus.

The chief localities where these multizoned skarns are developed and have been studied are as follows (pl. XXIV):

1. No. 1 Prospect, Kilchrist (loc. A, sub-locs. *a–g*, fig. 15).
2. At the contact of a dolomite island in the granite, 300 yards SE. of the old Manse, Kilchrist (loc. B).
3. In the quarries lying 350 yards ESE. of no. 1 Prospect, Kilchrist.

4. No 2 Prospect, Kilchrist, with the quarries adjacent (loc. C).
5. Kilbride (loc. D and loc. E).
6. Camas Malag (Loch Slapin), 1730 yards S. 13° W. of Kilbride (loc. G, and loc. F adjacent thereto).
7. South margin of the granite, on the west bank of the burn, tributary to the Allt Leth Slighe, 2030 yards S. 3° E. of Kilbride (loc. H).
8. South margin of the granite, N. of Allt Leth Slighe, 2100 yards S. 7° E. of Kilbride (loc. J).
9. South margin of the granite, small isolated lens of granite in limestone of embayment, 2380 yards S. 27° E. of Kilbride (loc. K).

In the preliminary note on the development of these dolomite skarns, a tabulation of their mineral content was given and the broad zonal distribution of these minerals indicated. An earlier paper (Tilley, 1947) had presented the characteristics of the mineral cuspidine, a prominent constituent, while in 1949 a typical zoned skarn from Kilchrist had been figured (Tilley, 1949).

The skarn zones are in general multiple, but of varying complexity. Broadly a twofold division can be recognized: (*a*) the primary zones, (*b*) the secondary zones, each in themselves multiple. The simplest skarn development is that unaccompanied by iron ores or a boron-fluorine pneumatolysis, illustrated in the contact skarns of the quarries near no. 2 Prospect, Kilchrist (loc. C), in the quarries SE. of no. 1 Prospect, Kilchrist, and on the south margin of the granite (loc. K). The succession seen in a skarn zone of the quarries 350 yards SE. of no. 1 Prospect, Kilchrist, illustrates this simpler development (fig. 1*a*).

At the contact with the granite a succession built of hedenbergite-plagioclase-wollastonite solid solution[1] is followed by a zone of white diopside (with accessory spinel and xanthophyllite), a thin band of forsterite which then gives place to the country-rock, a forsterite-bearing marble. Close to the contact the normal hornblende-biotite-granite gives place to a clinopyroxene-bearing granite usually enriched in plagioclase.

When, as exemplified in this section, wollastonite occurs in association with hedenbergite, the mineral is an iron-wollastonite. An analysed example from another skarn has been described (Tilley, 1948, *c*). In the development of this multizoned skarn there is evidence that both silica and alumina have been transferred from the granite; the migration of the latter is, however, largely restricted to the inner zones, those bearing plagioclase. Iron has also diffused outwards in the formation of the pyroxene and wollastonite.

[1] In place of this zone a grossular-wollastonite assemblage may be developed, as in the skarns of the quarries near no. 2 Prospect, Kilchrist.

The central zone of clinopyroxene is doubtless silicated dolomite. The forsterite zone resembles in its position the shell of forsterite which follows on diopside at the periphery of the 'sponge forms' in the main

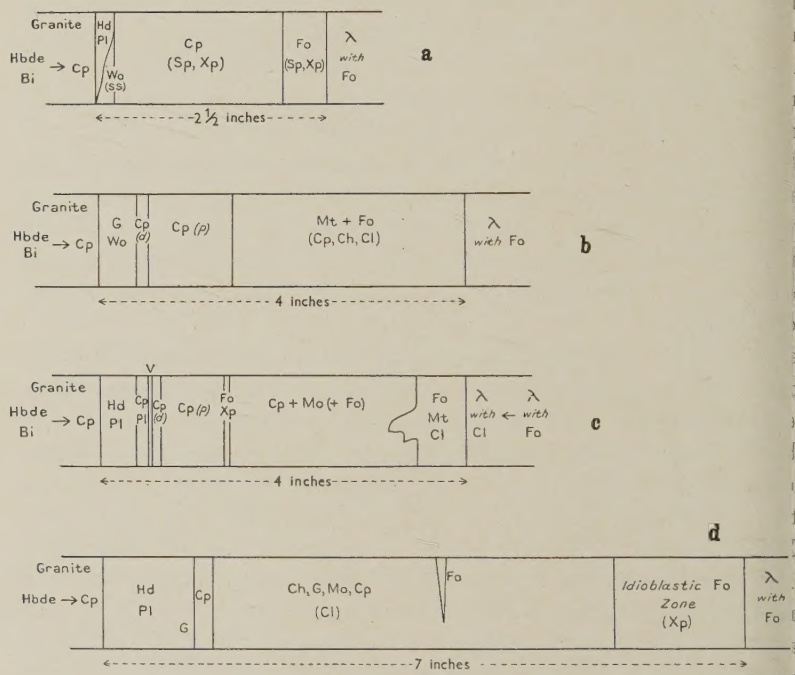


FIG. 1. Sections across zoned skarns, granite contacts.

(*a*) Quarries 350 yards SE. of no. 1 Prospect, Kilchrist. (*b*) No. 1 Prospect, Kilchrist, loc. A. (*c*) Island of dolomite in granite, 300 yards ESE. of the old Manse, Kilchrist, loc. B. (*d*) Kilbride, loc. E.

Hbde hornblende, Bi biotite, Cp clinopyroxene, Pl plagioclase, G grossular, Hd hedenbergite, Wo (Wo_{ss}) wollastonite and solid solutions, Ch chondrodite, Cl clinohumite, Mo monticellite, Fo forsterite, Xp xanthophyllite, Sp spinel, Mt magnetite, V idocrase, λ limestone (dolomite).

aureole. At locality K, as already noted (Tilley, 1949), the inner zone is built of grossular and wollastonite and the contiguous granite is of alkali type. Unlike the former example there has been no reciprocal transfer of material from the dolomite to the granite.

A more complicated zonal development is that characteristic of Kilbride (loc. E, fig. 1 *d*), where an inner hedenbergite-plagioclase assemblage (sometimes associated with iron-wollastonite) is followed by a narrow zone of clinopyroxene and then by a much wider zone built of

chondrodite, grossular, monticellite, and diopside±clinohumite, in which typically the monticellite occurs as large crystals enveloping diopside and the humite group minerals (pl. XX, fig. 6). Some parts of this particular zone are built essentially of chondrodite and grossular. This wide zone is followed by a remarkable zone built essentially of idioblastic forsterite, often with a minimum of interstitial calcite and some associated xanthophyllite. This zone gives place to the normal forsterite-marble.

Skarn zones which include layers containing both fluosilicates and magnetite are, however, the commonest met with. Those of no. 1 Prospect, Kilchrist, are typical, an example having been figured in an earlier paper (Tilley, 1949). Very characteristic here and at Kilbride is the succession from a magnesia-free grossular-wollastonite zone to two successive clinopyroxene zones, one composed almost wholly of green diopside, usually fine grained and the other, the next outward, of a pale diopside typically coarse grained. As exemplified in a Kilbride example the pyroxenes of both zones contain alumina (analyses 19 and 20 of table XIV). The pale clinopyroxene zone is followed by a zone of variable width forming the essential ore skarn, built of magnetite, and chondrodite, diopside, clinohumite, and forsterite in varying proportion (pl. XX, fig. 5). This zone abuts against the normal forsterite-marble.

A still more complicated multizoned skarn occurs at loc. B, at the contact of the island of dolomite. Here the inner zone is a hedenbergite-plagioclase assemblage, followed by a lighter coloured diopside-plagioclase aggregate.

The usual dark and pale pyroxene zones with an inconstant forsterite layer are succeeded by a zone built of monticellite, clinopyroxene, and forsterite, sometimes with magnetite, and this again by a magnetite-forsterite zone (ore skarn) which shows irregular contacts with it, and which is in part clearly replacing it, as veins of this assemblage penetrate through the monticellite-bearing assemblage (fig. 1 *c*). Both the monticellite and forsterite-magnetite zones may carry subordinate fluosilicates of the humite group. Against the latter the dolomite carries clinohumite, eventually giving way to the normal forsterite type.

The skarn zone of greatest complexity is that met with at Kilbride (loc. D *b*) in a small gully west of the ludwigite locality (loc. D *a*). Here the skarn zones attain a wider dimension. The succession is effectively that portrayed in fig. 2. The distinguishing features of this skarn development are the presence of two monticellite zones separated by a broad zone rich in diopside and cuspidine, but also carrying one or both

of the fluosilicates, clinohumite, and less frequently chondrodite. Each monticellite zone is associated with a narrow clinohumite zone.

The surrounding marble is a periclase-spinel-forsterite type, which in the vicinity of the skarn zones develops clinohumite as idioblastic crystals and as polycrystalline aggregates taking the form of the rounded

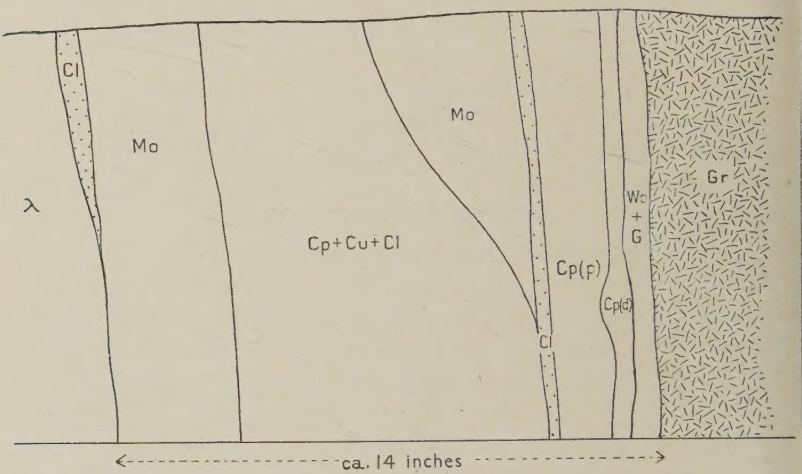


FIG. 2. Section across zoned skarns, Kilbride (loc. D *b*).

Gr granite with clinopyroxene, Wo+G wollastonite-grossular zone, Cp(d) dark clinopyroxene zone, Cp(p) pale clinopyroxene zone, Cl clinohumite zone, Mo monticellite zone, Cp+Cu+Cl clinopyroxene-cuspidine-clinohumite zone, λ marble with clinohumite.

original, but now transformed, forsterite grains. Periclase disappears close to the contact, and accessory fluoborite and ludwigite, and venules of fluorite may appear in the marble. A photograph of this outer region of the skarn succession and the associated marble is given in fig. 3.

A few yards east of this skarn exposure is the outcrop of the skarn assemblage yielding ludwigite in some abundance. The exposures as now seen occur on the faces of a pit excavated along the granite contact. The limestone horizon appears to be that of the Strath Suardal zone, for ring structures are seen in the outcrops of limestone adjacent and the prevailing type of contact metamorphism is that of forsterite and brucite marbles. The principal skarn from this locality is a ludwigite-chrondrodite-magnetite skarn in which fluoborite and szaibelyite may appear as accessories. The granite in contact with the ore skarn in the pit is a porphyritic type of hornblende-granophyre showing phenocrysts

quartz, microperthite, oligoclase, and brown-green hornblende in a artz-microperthitic base, partly in micrographic intergrowth.

The ludwigite skarn is dull black on the weathered surface, frequently nded, due to thin impersistent streaks of white or creamy white ondrodite. On the fractured surface the ludwigite appears as radiate sters of velvet black needles reaching to 1 mm. in length. Magnetite

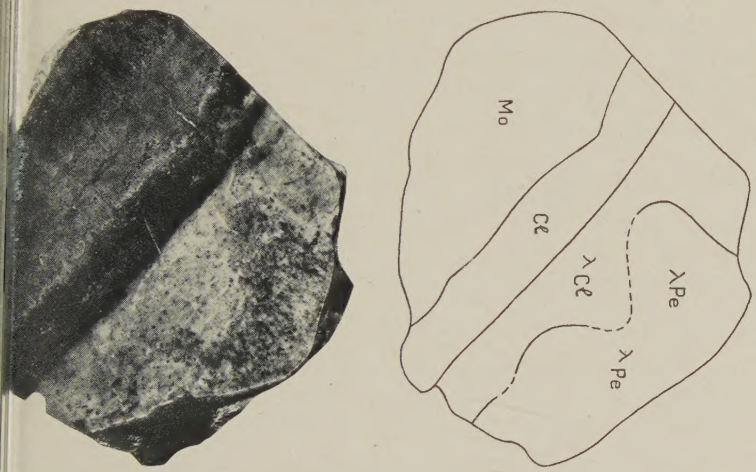


FIG. 3. Zoned skarn at granite contact, Kilbride (loc. D). $\times\frac{1}{2}$.

Edge against marble: marble with periclase λPe, forsterite, and spinel, passing rwards into periclase-free marble with forsterite, clinohumite, spinel, some dwigite λCl; contact against clinohumite zone (Cl) and broad monticellite zone Mo). The marble at the contact against the clinohumite zone is rich in idioblastic inohumite and has forsterite grains in large part replaced by a polycrystalline ggregate of clinohumite; there are minute sprays of ludwigite fibres. The photo-raph shows the clearer zones of the marble against the clinohumite zone, freed om dark periclase. In the outer zone of the marble periclase, brown, is mantled y brucite bearing minute iron-ore grains, the ferropericlase yielding brucite with paration of iron oxide. In the parts of the marble bearing both clinohumite and rsterite the former, if as single grains, is idioblastic, forsterite as rounded grains, en often replaced by aggregates of clinohumite.

usually present, though in variable amount, and some of the skarns ntain sporadic patches of basic copper carbonates along joints. Less mmonly a skarn composed of diopside and magnetite with accessory rossular and idocrase is found.

Two other skarn localities along the granite edge occur, one to the st and the other to the west at the angle of the contact where a pro-usion of the granite is directed northward. It is probably this latter

locality to which A. Geikie refers in his comment on magnetite a Kilbride (Ancient volcanoes, 1897, vol. 2, p. 384). Magnetite is her accompanied by amphibole (altered in part to talc) and clinochlore, an the associated marbles carry idocrase and grossular. Fluorite and blend are also present. The important skarn occurrence on the west contac of the protrusion of granite at Kilbride (loc. E) has already been men tion and will be discussed in more detail in the sequel.

At the western end of the granite the most important development o skarns is that at Camas Malag on Loch Slapin. Here within an embay ment of dolomite three localities are concerned (loc. F, loc. G *a*, and loc

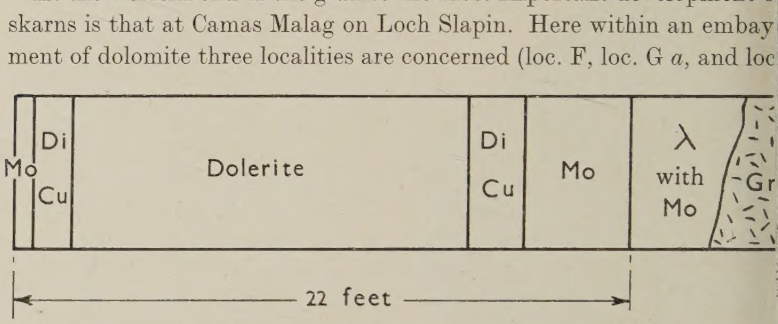


FIG. 4. Skarn development at edges of a pre-granitic dolerite dike at Cama Malag (loc. G *a*): a diopside(Di)-cuspidine(Cu)-magnetite skarn zone is followed outwards by a monticellite(Mo)-magnetite skarn, λ limestone, Gr granite.

G *b*). It is these skarns which have provided the most extensive develop ment of monticellite and cuspidine in the region, and have provided at loc. G *b* the special mineral assemblage carrying the new boron mineral harkerite. The position of these skarn localities is indicated on the map (pl. XXIV), and in the accompanying section (fig. 4) the zonal arrange ment at loc. G *a* is set out.

A prominent dolerite dike (15 feet wide) seen in the burn section Allt na Garbhlain is sheathed by skarns built of diopside-cuspidine-magnetite ±chondrodite and monticellite±magnetite, while there is a strong development of monticellite-marbles near the adjacent granite contact (fig. 4). In the diopside-cuspidine-magnetite zone occur 'sponge forms' with their own suite of skarn successions. It is clear that the contact surfaces of this pre-granitic dolerite in the marble have provided channel ways for the introduction of the skarn producing solutions.

At locality F at the end of the embayment an important group of skarns carrying idioblastic cuspidine, forsterite, and clinohumite is developed, and the associated marble carries small idioblastic crystals of clinohumite. Here occurs the clinopyroxene-fluorite skarn, of which an analysis is given in table XIII, no. 15.

The locality of the harkerite-bearing skarns (loc. G *b*) is at the granite ntact about 50 yards SE. of the chief skarn locality in the burn section. ne harkerite occurs in a band a foot or two in length and 6 8 inches thickness in the dolomite associated with monticellite to form a arkerite-monticellite skarn. While monticellite is its commonest asso- ate, other minerals found with it include diopside, calcite, and the es magnetite, bornite, chalcosine, and blende.

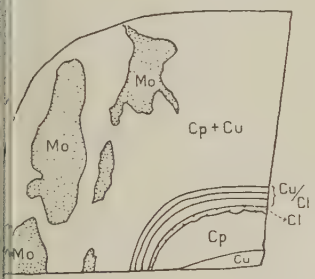


Fig. 5 ($\times \frac{3}{5}$).

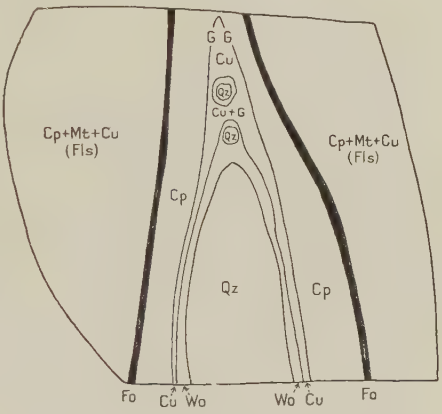


Fig. 6 ($\times \frac{3}{4}$).

Zoned skarns around 'sponge forms' near granite contact, Camas Malag, Loch lapin.

The complicated zonal succession is well displayed. In fig. 5 the 'sponge form' self is wholly replaced by cuspidine and clinopyroxene and is succeeded by lternating narrow bands of cuspidine and clinohumite (Cu/Cl). In fig. 6 the zones how bilateral symmetry across a chert core.

Cp clinopyroxene, Mo monticellite, Qz quartz, Wo wollastonite, Fo forsterite, grossular, Fls chondrodite-clinohumite, Mt magnetite.

The humite minerals, clinohumite and chrondrodite, are found in the ssociated dolomite, and clinohumite may appear as a thin band in direct contact with the harkerite skarn. Further data on the harkerite paragenesis are given in the section on the mineralogy of the skarns (p. 640).

Three other localities of skarns on the south edge of the granite are marked on the map. They are localities H, J, and K, the first the occurrence of a garnetized dolerite dike carrying idocrase, the second a magnetite skarn, and the last surrounding the lens of alkali-granite previously described.

3. Skarn successions associated with chert nodules ('sponge forms').

The 'sponge forms' of the Strath Suardal dolomite when involved in the skarn zones exhibit their own succession of zones and have suffered fundamental metasomatic change. The best examples come from Kilchrist (loc. A) and Camas Malag (loc. G *a*). In the aureole we have noted that in the highest grades of metamorphism the core of the nodules may be

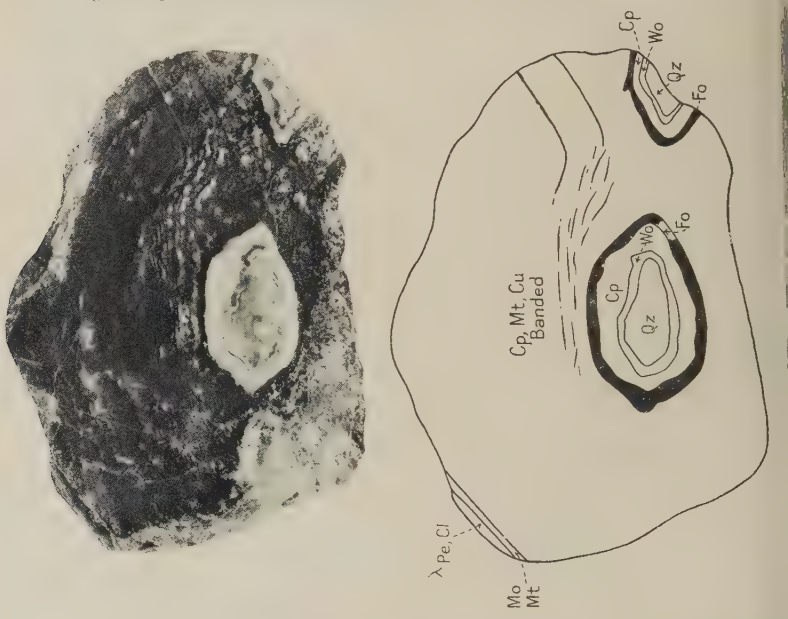


Fig. 7. Zoned skarn around 'sponge form', Camas Malag (loc. G), Loch Slapin. $\times\frac{2}{5}$.

The chert core is surrounded successively by a zone of wollastonite, clinopyroxene, fine-grained forsterite, with some chondrodite. The dark groundmass beyond is ore skarn built of clinopyroxene, magnetite, cuspidine, and subordinate fluosilicate (chondrodite series), often banded with layers rich in magnetite. At the periphery (bottom left) the ore skarn is followed by a zone of magnetite-monticellite with some calcite passing to marble with periclase, spinel, and some clinohumite.

replaced partly or wholly by diopside. In the skarn zones where the core is transformed by metasomatism the ultimate replacement is by wollastonite, cuspidine, or sometimes even by apophyllite. In the larger nodules a core of recrystallized quartz is preserved, but it may be traversed by venules of wollastonite, diopside, pectolite, or xonotlite.

A typical succession at Kilchrist outward from the quartz core is wollastonite as radiating pencils at right angles to the core surface, followed by diopside or diopside and cuspidine with grossular or idocrase, eventually giving place to a magnetite-cuspidine skarn. In some examples the core of quartz is partially or wholly replaced by apophyllite, optically positive, and showing anomalous brown interference tints.

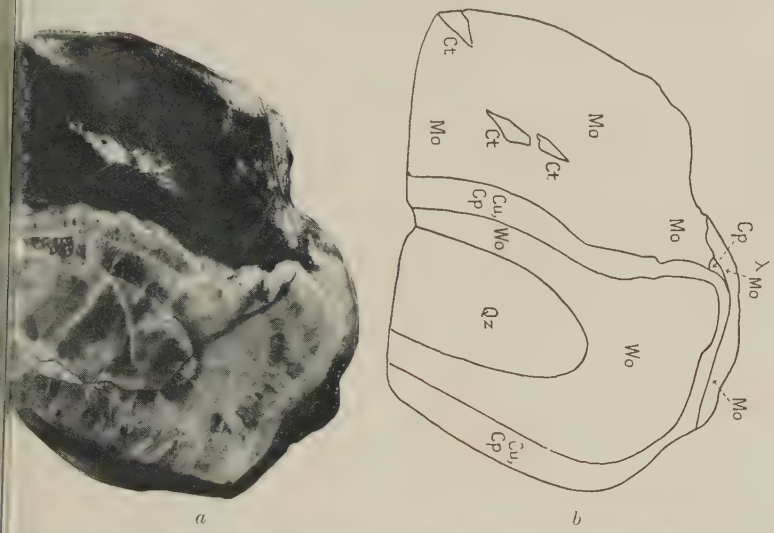


FIG. 8. Zoned skarn around 'sponge form' close to granite, Camas Malag (loc. G), Loch Slapin. $\times\frac{4}{5}$.

A recrystallized quartz core (with some diopside, wollastonite, and pectolite) is surrounded by a broad wollastonite zone (pencils at right angles to contact). Then follows a narrow zone of cuspidine which contains some intergrown clinopyroxene and calcite. This zone borders a marble with monticellite (right) and at the top a broad zone of monticellite skarn containing patches of calcite.

Both at Kilchrist and at Camas Malag the border between the quartz core and the radiate wollastonite zone may be occupied by a very narrow band less than $\frac{3}{4}$ mm. in thickness of a colourless mineral in the form of fibres of parallel orientation set parallel to the wollastonite pencils. This mineral has a refraction, α 1·620, γ 1·628, a little less than that of the wollastonite, elongation positive, sensibly uniaxial, optically positive. It appears to be a hydrated calcium silicate, though its properties do not accord with any of the described species of that type of composition. When a slice is strongly heated, the band containing this mineral becomes turbid and eventually appears to be converted to wollastonite, losing

then something of its parallel orientation. The exact nature of this material, owing to the scarcity of it, has not yet been established.

The 'sponge forms' in the skarn zones of Camas Malag show more fundamental changes and a more complicated zonal succession outward from the core. These transformations are illustrated in figs. 5–8. A wollastonite sheath may be followed by a zone of diopside, a zone built of forsterite with which chondrodite or clinohumite may be associated, that zone being succeeded by an ore skarn often in a banded succession, Eozoon-like, of diopside, cuspidine, and clinohumite with magnetite (fig. 7). An important variant of this succession is seen in the examples of figs. 6 and 8, where the wollastonite zone is succeeded by a zone of cuspidine with granules of diopside and calcite, in which the cuspidine has a similar radiate texture as the wollastonite but so oriented that the lath-like sections show negative elongation, suggesting that the cuspidine is developed flattened parallel to (001). This cuspidine zone may be followed by a zone of brown monticellite (fig. 8) or a diopside zone (fig. 6). In the smaller 'sponge forms' the core may be replaced at its centre by wollastonite or by cuspidine (fig. 5).

TABLE I. Minerals present in aureole and skarn zones.

	Skarn zones.		
Aureole beyond the skarn zones.	Group I, Primary skarns.	Group II, Boron-fluorine ore skarns.	
Talc	Grossular-andradite*	Magnetite*	Grossular-andradite
Tremolite	Wollastonite* solid solutions	Tremolite	Hydrogrossular
Forsterite	Diopside-hedenbergite*	Forsterite*	Idocrase
Diopside	Spinel	Diopside*	Bornite
Periclase	Plagioclase*	Monticellite*	Chalcosine
Wollastonite (rare)	Idocrase	Cuspidine*	Covelline
Spinel	Xanthophyllite	Fluorite	Chalcopyrite
Idocrase	Phlogopite	Chondrodite*	Pyrite
Grossular	Orthite	Humite	Blende
Phlogopite	Clinozoisite-epidote	Clinohumite*	Galena
Brucite	Prehnite	Ludwigite	Chessylite
Serpentine	Apophyllite	Fluoborite	Malachite
Chlorite	Pectolite	Szaibelyite	
Hydromagnesite	Xonotlite	Datolite	
		Harkerite	

* Most abundant skarn minerals.

4. MINERALOGY OF THE SKARN ZONES.

The mineral development in the aureole and in the skarn zones of the granite is set out in table I, the columns separating the minerals

aracteristic of (*a*) the aureole beyond the skarn zones, (*b*) the primary arns (group I), and (*c*) the secondary (group II) skarns.

Clinopyroxenes.

Clinopyroxenes play a prominent role both in the skarns of groups I d II. Analyses of the clinopyroxenes of the twin zones of group I at lbride (loc. D *b*) are given in table II together with their optical perties. Comment on the character of these zones will be made in sequel (p. 646). An analysis of a skarn zone from Camas Malag, lt of clinopyroxene and fluorite enables us to calculate the com-sition of a more ferriferous diopside in a skarn zone. The analysis is en in table III.

Hedenbergite is not an uncommon constituent of the zone which kes the place of the grossular-wollastonite assemblages at some of the nite contacts. A typical example is provided in the hedenbergite-agioclase skarn of Kilbride (loc. E). Here the hedenbergite is prac-ally free of magnesia (analysis 11 of table XIII). It is with the more

Table II. Analyses of diopside.

	1.	Metal atoms to 6 oxygens.		2.	Metal atoms to 6 oxygens.	
O_2	50·74	1·876	2·00	49·81	1·828	2·00
$_2O_3$	3·67	0·159	0·124	6·42	0·277	0·172
			0·035			0·105
$_2O_3$	0·96	0·026		1·25	0·035	
O	0·79	0·024	1·02	1·03	0·030	1·01
nO	0·08	—		0·06	—	
gO	16·92	0·938		15·22	0·837	
O	25·78	1·009		25·92	1·011	
O_2	0·11	—		0·14	—	
O—	0·11			0·06		
O+	0·54	α 1·669		0·13	α 1·680	
$_2$	0·24	γ 1·695		0·15	γ 1·704	
	99·94			100·19		

1. Diopside of pale clinopyroxene zone
2. Diopside of dark clinopyroxene zone
} Skarn, Kilbride, loc. D *b*.

denbergite pyroxenes that wollastonites, having the properties of on-wollastonite, occur. An example from Camas Malag has already en described (Tilley, 1948, *c*). At Kilbride (loc. E) pockets of an iron-llastonite are associated with the hedenbergite-plagioclase skarn

TABLE III. Analysis of clinopyroxene-fluorite skarn.

	1.	2.	Metal atoms to 6 oxygens.	
SiO_2	38·48	51·08	1·921	2·00
Al_2O_3	1·99	2·64	0·110 (0·08 / 0·03)	
Fe_2O_3	0·44	0·58	0·02	1·029
FeO	5·78	7·67	0·248	
MnO	0·15	0·20		
MgO	9·77	12·97	0·731	
CaO	35·59	24·37	0·982	2·01
Na_2O	0·05	0·07		
K_2O	0·02	0·03		
H_2O-	0·16	0·21		
H_2O+	nil	—		
TiO_2	0·08	0·10		
P_2O_5	0·05	0·07	α 1·682, γ 1·710, γ:c=40°,	
SO_3	0·38		$2V_\gamma$ 60°	
F	11·52	—		
	104·46	99·99		
Less O=F ...	4·85			
	99·61			

1. Clinopyroxene-fluorite skarn, Camas Malag (loc. F). Diopside 75·5, fluorite 23·6, calcium sulphate 0·6 %.
2. Clinopyroxene, after deducting fluorite, and SO_3 as calcium sulphate.

noted above. It shows γ 1·655, $\alpha' : c = 44°$, in sections perpendicular to the zone of cleavages, thus corresponding to a composition $CaSiO_3$ 80 $FeSiO_3$ 20 wt. % (Plate XXI, fig. 3).

In association with diopside pyroxene in the skarn zones—for example those occurring as borders to 'sponge forms', or in the grossular-wollastonite assemblages at the granite contacts—the wollastonite shows the normal optic properties of the iron-free mineral.

Fluorine-bearing minerals.

Humite series.—The chief minerals of the humite series developed in the skarns are chondrodite and clinohumite. Humite is much less common, while norbergite has not been recorded. Both chondrodite and clinohumite are developed at the chief localities; humite is present in parallel growth with clinohumite at Kilbride in the clinohumite skarn of loc. D *b* and with chondrodite at no. 1 Prospect, Kilchrist. Parallel growth of both with forsterite are also recorded. A fine example of clinohumite-forsterite intergrowth is figured from the ore skarn of

ilchrist (pl. XX, figs. 1 and 2). Repeated twinning on (001) is characteristic of both fluosilicates: in addition, twinning on (105) or ($\bar{3}05$) occur in chondrodite and on ($\bar{1}03$) in clinohumite (pl. XX, fig. 4).

Both minerals occur as idioblastic crystals in the marbles adjacent to the ore skarns. At Kilbride and Camas Malag both minerals form polycrystalline aggregates replacing the rounded forms of forsterite in zones in the marble adjacent to the skarns (pl. XVIII, fig. 2). In the monticellite bands of Kilbride (loc. D *b*) the monticellite close to the fluosilicate bands frequently contains clinohumite enclosures. In some cases the relation is found to be that of parallel growths as illustrated in pl. XX, fig. 3. The optical relations of the oriented intergrowths found among the olivine and humite-group minerals in the skarns are depicted in text-fig. 9.

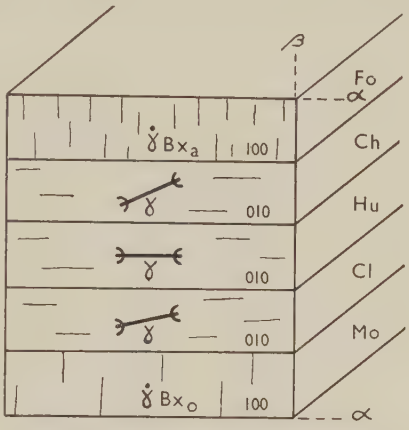


Fig. 9. Relations of parallel growths of the humite-group minerals (Ch, Hu, Cl), forsterite (Fo), and monticellite (Mo) in the skarns of the contact zone of the Beinn an Dubhaich granite.

An analysis of a clinohumite skarn forming a half-inch band at the outer edge of the multizoned skarn of loc. D *b* Kilbride is given in table XIII, no. 12. It contains some intergrown humite and has as accessories, calcite, clinochlore, xanthophyllite, fluorite, and a little magnetite.

Cuspidine.—The characteristic properties of this important mineral of the skarns have already been described (Tilley, 1947). Magnificent specimens of idioblastic cuspidine are recorded from Camas Malag, especially at locality F and illustrated in pl. XIX, figs. 1 and 2.

Fluoborite is richly present in a number of the ludwigite skarns, but also appears as an accessory in some of the magnetite-chondrodite assemblages. It is intimately associated with chondrodite in the form of hexagonal prisms. Examples studied show ω 1·550, ϵ 1·522, corresponding approximately to a composition of 50 % of the hydroxyl end-member of the series $3MgO.B_2O_3.3Mg(OH,F)_2$ (Schaller, 1942). A specimen from a ludwigite skarn gave ω 1·556, ϵ 1·528. A section of a fluoborite-bearing skarn is figured in pl. XIX, fig. 5.

Boron-bearing minerals.

In addition to fluoborite, szaibelyite, datolite, ludwigite, and the new boron mineral harkerite call for some notice.

Szaibelyite ($2MgO.B_2O_3.H_2O$) occurs in magnetite skarns at no. 1 Prospect, Kilchrist, and at Kilbride. At the former locality the rock is a magnetite-ludwigite-chondrodite-carbonate assemblage. Set in the carbonate is fine-granular magnetite and coarser chondrodite, but there are knots built of an intergrowth of magnetite and lath-like ludwigite in part coverted to haematite. At the periphery of these knots is szaibelyite as a mica-like development of matted fibres or plates with high birefringence, negative elongation, pseudo-uniaxial, negative sign, α 1·575, γ 1·650. In its occurrence the Skye szaibelyite resembles that of Tallgruvan of the Norberg district, Sweden, described by Geijer (1927, 1939).

Datolite is recorded at no. 1 Prospect, Kilchrist, at the contact of the grossular-wollastonite skarn with the alkali type of granite there produced. It appears replacing quartz of the granite at the immediate contact and is usually accompanied by prehnite which may similarly replace the adjacent orthoclase. As spherulitic botryolite it occurs in a similar environment in venules at the contact. A secondary mineral entirely, it derives its boron from solutions which have given rise to the associated ore skarns.

Ludwigite.—A description of the chief occurrence of ludwigite has already been given (p. 628). Besides its development at Kilbride, it is found as an accessory in the skarns of Kilchrist and in the marbles of Kilbride and Camas Malag. In its chief occurrence it is a ferroludwigite, transparent only in the thinnest fibres; then pleochroic, γ brown, α and β green, absorption $\gamma > \alpha = \beta$. Straight extinction, γ parallel to the fibre length. The mineral is magnetic. An analysis of the Kilbride ludwigite is given in table IV. After minor amounts of chondrodite and magnetite have been removed, this analysis shows close agreement with that of the end-member ferroludwigite (col. 2).

An analysis of a typical ludwigite assemblage at Kilbride, a ludwigite-chondrodite-magnetite skarn, is given in the list of analyses (table XIII, no. 14). It is quoted below (table V).

The ludwigite paragenesis with chondrodite (or clinohumite) and fluoborite characterizes other ludwigite occurrences such as those of the Norberg district (Geijer, 1927, 1939), Rézbánya (Watanabe, 1939), and the Suian district of Korea (Watanabe, 1939, 1943). This type of paragenesis may be conveniently illustrated graphically as in the tetrahedron

TABLE IV. Analysis of ludwigite.

	1.	2.	Mol. ratios.	Chondrodite.	Magnetite.	Remainder.	
O_2	1·28	—	21	21	—		
l_2O_3 ...	2·25	—	22	—	—	226	1·0
e_2O_3... ...	35·31	37·95	221	—	17		
iO_2	0·30	—	4	—	—	218	0·97
eO	16·36	17·08	227	—	17		
nO	0·29	—	4	—	—		
gO	28·33	28·46	708	52	—	656	2·90
aO	0·07	—	—	—	—		
$_2O_3$	15·74	16·51	226	—	—	226	1·0
$_2O-$ ...	0·19				—		
$_2O+$ ...	0·44	—	—	—	—		
	100·56	100·00					

1. Ludwigite, loc. D *a*, Kilbride, from a ludwigite-chondrodite skarn with magnetite.

2. Ferroludwigite (theoretical, $3MgO.B_2O_3.FeO.Fe_2O_3$).

TABLE V. Analysis of ludwigite-chrondrodite-magnetite skarn.

	Mol. ratios.	Chondrodite.	Magnetite.	Remainder.		
iO_2	5·47	91	91	—		
l_2O_3... ...	2·62	25	—	—	166	1·0
e_2O_3... ...	33·29	208	—	67		
iO_2	0·11	1	—	—		
eO	13·78	191	—	67	129	0·77
nO	0·30	4	—	—		0·99
gO	30·20	755	227	—	534	0·22
aO	0·36	6	—	—		3·00
$_2O_5$	0·01	—	—	—		
$_2O_3$	11·53	166	—	—	166	1·0
$_2O-$ ...	0·09	—				
$_2O+$ ...	2·47	—				
... ...	0·34	—				
	100·57 less 0·14 O=F, 100·43					

The composition of this rock (sp. gr. 3·515) is estimated as ludwigite 67·1, chondrodite 15·5, magnetite 15·5 wt. %. This type of skarn is figured in pl. XIX, fig. 6.

figure (fig. 10) where the apices of the tetrahedron are represented by MgF_2, forsterite, kotoite ($3MgO.B_2O_3$), and magnetite. The compositions of fluoborite, the humite minerals (represented by chondrodite or clinohumite), and ludwigite are represented on the tetrahedron edges.

The four-phase assemblage of the Skye boron-fluorine minerals is represented by the field ludwigite-chondrodite-magnetite-fluoborite. Other possible four-phase assemblages are also indicated.

Kotoite has not been recorded in the Skye skarns, nor should it be expected in ferroludwigite associations with magnetite, for we may write kotoite + magnetite = ludwigite, and further forsterite + fluoborite = chondrodite (clinohumite) + kotoite, so that forsterite-fluoborite associations are further unlikely. The four-phase assemblages carrying kotoite have been recorded from the type locality for kotoite in Korea.

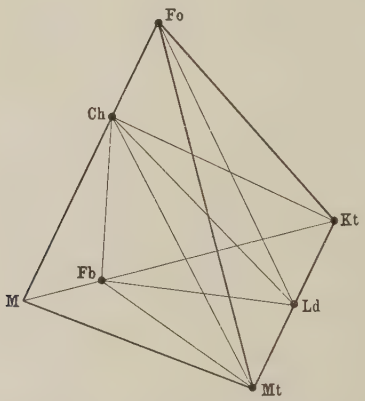


FIG. 10. Four-phase assemblages in the system Mg_2SiO_4 (Fo)–MgF_2 (M)–Fe_3O_4 (Mt)–$3MgO.B_2O_3$ (Kt kotoite). Ch chondrodite, Fb fluoborite, Ld ludwigite.

Harkerite.—This mineral is developed at one locality only, that of Camas Malag (G *b*). It occurs typically in the form of simple octahedra (pl. XXI, fig. 2). When fresh it is colourless with a vitreous lustre, but its ready alteration to fine calcite leads to its appearance as white masses in which, however, the octahedral outlines can still be observed on fractured surfaces. Bands of the harkerite-monticellite skarn may pass into zones of almost pure harkerite-rock in which, however, incipient alteration to calcite is seldom absent. This alteration of the mineral may affect the core of the mineral or be confined to narrow zones within the crystal following the outline of the idioblasts. The zoned character of the crystals as seen under a hand-lens is due to this characteristic feature of the mineral.

In section harkerite is colourless with a refractive index of 1·653. It is

us closely similar in refraction to its chief associate, monticellite. In is environment the mineral is distinguished in ordinary light by its ioblastic development in contrast to the granular character of the osaic of monticellite, irregular crystals of which may also be enclosed ithin the harkerite. The mineral shows no signs of a significant cleavage d has a density of 2·959 at 20° C. Many of its sections behave as a bic mineral, being isotropic, but frequently an anomalous birefringence met with resembling in this regard the grossular garnet of contact nes.

The optical structure corresponds closely to that of the Elba type of rnet in which the octahedral crystal behaves as if built of pyramidal its with their bases in the octahedron face. Thus in sections parallel (100) the squarish sections extinguish in the parallel position and parate into four fields with weak birefringence in the diagonal position ving a low first-order grey interference tint with each of the fields otically negative, the fast ray being normal to the edge length. Sections t parallel to the octahedron face show an isotropic, triangular core ith the peripheral sectors weakly birefringent and fast ray across their ng edges. This structure, however, does not always involve the whole ystal, for it may be intercepted by, or alternate with, narrow zones llowing the outline which remain isotropic, a feature recognized in me grossulars.

The mineral has been isolated from a skarn containing accessory ore inerals, bornite, chalcosine, and magnetite, together with a little iopside. From these minerals it is readily separated by heavy liquids, ut the removal of the intimately associated fine-grained calcite occurng in zones of the harkerite and in minute cracks has proved a more ifficult operation. The uncontaminated mineral dissolves with effervescence in acetic or hydrochloric acid and gives a good flame reaction or boron. Heated to 850° C. for long periods, harkerite breaks down to a urbid brown sheaf-like product spreading through it in spherulitic ashion. After 30 hours' heating some of the original harkerite is still reserved as islands in this transformation product. After ten days' eating at 850° C., followed by five days at 950° C., this sheaf-like roduct had itself broken down to yield in slice a dense grained material lmost isotropic.

An analysis of harkerite carried out by Mr. H. C. G. Vincent is set own in table VI. This analysis shows that harkerite is of complex omposition and may be described as essentially a carbonate-borosilicate f calcium and magnesium containing chlorine.

TABLE VI. Analysis of harkerite.

	1.	2.	3.	4.	5.	6.	7.	
	wt. %	Mol. ratios.	Metals.	O,OH, Cl.	Metals to 156 (O,OH, Cl).	Calc. wt. %	Metals to 156 (O,OH, Cl).	
SiO_2 ...	14·17	0·2359	0·2359	0·4718	13·35	13·67	13	24
B_2O_3 ...	7·77	0·1116	0·2232	0·3348	12·63	6·71	11	
Al_2O_3 ...	2·84	0·0279	0·0558	0·0837	3·16	3·12	3·5	20
Fe_2O_3 ...	0·85	0·0053	0·0106	0·0159	0·60	0·70	0·5	
FeO ...	0·46	0·0064	0·0064	0·0064	0·36	0·63	0·5	
MnO ...	0·02	—	—	—	—	—	—	
MgO ...	11·15	0·2765	0·2765	0·2765	15·65	10·87	15·5	
CaO ...	46·23	0·8243	0·8243	0·8243	46·65	47·12	48	
CO_2 ...	14·94	0·3394	0·3394	0·6788	19·21	15·42	20	
Cl	1·36	0·0383	—	0·0383	2·17	1·25	2	
H_2O+ ...	0·81	0·0450	0·0900	0·0450	5·09	0·79	5	
H_2O- ...	0·11	—	—	—	—	—	—	
	100·71			2·7755		100·28		
Less O = Cl ...	0·31			0·0191		0·28		
	100·40			2·7564		100·00		

For the following X-ray data on the mineral I am indebted to Dr. N. F. M. Henry. Oscillation, Laue, and powder photographs, the last with both iron and copper radiation, were taken. The cell side was found to be 29·53 Å. to an accuracy of 1 part in 3000. The odd-layer lines in the oscillation photographs were very weak, thus showing the existence of a pseudo-repeat distance of 14·76, i.e. half the true repeat distance. The Laue group was found to be $m3m$, and the results of tests for pyro- and piezo-electricity were both negative. Consequently it is very probable that the crystal class is $m3m$ (cubic holosymmetric). The large cell size makes the assignment of indices to the powder lines rather difficult. Even with a 19-cm. diameter camera and iron radiation it was not found possible to do this completely. With iron radiation seventy-three lines could be measured, while the use of copper radiation allows another nineteen to appear. The spacings and estimated relative intensities are given below, the four strongest lines being marked (table VII).

Taking a pseudo-cell as half 29·53 − 14·76 A. the molecular weight is 5731, and from this value the chemical analysis gives 157 as the number of anions in this pseudo-cell. In table VI the analysis on a basis of 156(O,OH,Cl) gives the figures set out in column 5, but these results manifestly do not provide the answer to the formula of this mineral.

TABLE VII. X-ray powder data for harkerite in Å.

No. of line.	*d*.	*I*.	No. of line.	*d*.	*I*.	No. of line.	*d*.	*I*.
1	5·22	s	26	2·22	vw	51	1·33	vw
2	4·45	w	27	2·16	vw	52	1·30	m
3	4·17	vw	28	2·13	s (3)	53	1·28	w
4	3·84	vw	29	2·07	m	54	1·26	w
5	3·69	vw	30	2·03	vw	55	1·25	w
6	3·61	vw	31	2·00	vw	56	1·22	w
7	3·58	vw	32	1·97	m	57	1·20	vw
8	3·39	m	33	1·92	w	58	1·18	vw
9	3·30	vw	34	1·84	s (2)	59	1·17	m
10	3·24	w	35	1·82	vw	60	1·15	vw
11	3·15	vw	36	1·80	vw	61	1·14	vw
12	3·01	m	37	1·74	m	62	1·13	vw
13	2·97	vw	38	1·70	m	63	1·11	w
14	2·95	vw	39	1·65	m	64	1·10	vw
15	2·84	m	40	1·61	w	65	1·08	vw
16	2·81	vw	41	1·60	vw	66	1·07	vw
17	2·66	w	42	1·57	w	67	1.06	w
18	2·61	vs (1)	43	1·55	vw	68	1.05	vw
19	2·53	vw	44	1·51	s (4)	69	1·0	w
20	2·50	vw	45	1·48	vw	70	1·03	vw
21	2·46	w	46	1·44	w	71	1·00	vw
22	2·39	vw	47	1·42	w	72	0·997	vw
23	2·37	vw	48	1·39	vw	73	0·986	m
24	2·33	w	49	1·38	vw			
25	2·25	vw	50	1·34	vw			

vs very strong, s strong, m medium, w weak, vw very weak. The four strongest lines are marked (1)–(4).

In columns 6 and 7 (table VI) are set down a calculated composition based on a pseudo-cell containing 156 anions with boron and silicon grouped together as a $(B,Si)O_4$ unit. The corresponding formula for this composition (columns 6 and 7) may be written as follows:

$$20CaCO_3.Ca_{28}(Mg_{15\cdot5}, Al_{3\cdot5}, Fe'''_{0\cdot5}, Fe''_{0\cdot5})_{[=20]}(B_{11},Si_{13})(O,OH,Cl)_{96}.$$

Comparison of columns 1 and 6 show greatest departure for CaO and B_2O_3. It is clear that harkerite calls for a detailed X-ray structural analysis, and a further chemical analysis of selected material is much to be desired. It is hoped that this further study may shortly be continued. Meantime the presentation of optical, chemical, and X-ray powder data may at this stage provide the means of ready recognition of the mineral, should it appear in other contact zones subject to boron metasomatism.

Monticellite.—This mineral is one of the more important silicates both of the ore-free and ore-bearing skarns of group II. In the marbles immediately adjacent to these skarns, brown idioblastic monticellite

with ϵ (111), m (110), h (011) prominent, may appear. The best examples of such monticellite-marbles occur at Camas Malag (loc. G *a*), where they are developed in association with ore skarns built essentially of monticellite and magnetite. The analysis of monticellite isolated from such a marble is quoted below. This monticellite shows close agreement in composition and properties with a Crestmore monticellite (col. 2) described by Moehlman and Gonyer (1934).

TABLE VIII. Analyses of monticellite. 1, Skye; 2, California.

	1.*	2.	Mol. ratios (1).	Magnetite.	Metals to 4 oxygens.	
SiO_2	36·55	37·46	609	—	0·990	
Al_2O_3	0·18	—	—	—	—	
Fe_2O_3	0·59	—	4	4	—	
FeO	4·71	3·98	65	4	0·099	0·998
MnO	—	0·52	—	—	—	
MgO	22·11	22·78	553	—	0·899	
CaO	35·21	35·20	628	—	1·021	
H_2O+	0·84	0·15				
H_2O-	0·56					
CO_2	nil					
	100·75	100·09				
Sp. gr.	3·09	3·083				
α	1·645	1·646				
β	1·654	1·653				
γ	1·660	1·660				
$2V_\alpha$	80°	78°				

* Corresponding to $CaMgSiO_4$ 88, $CaFeSiO_4$ 12 wt. %.

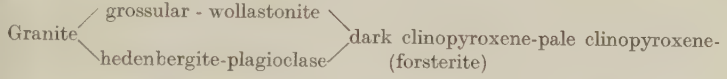
In the monticellite-magnetite skarns, while the monticellite is usually light brown in colour, there are areas of this mineral which are grey-white, and these tend to be free from magnetite—a clear indication that the skarn-forming solutions varied in their iron content yielding a more magnesian monticellite where magnetite was not precipitated in the immediate environment of the grey-white crystals. The monticellite of the skarns shows usually a poor cleavage and is usually in the form of simple crystals. In some of the skarns, however, a twinning on (031) has been recognized. Idioblastic crystals set in a matrix of chalcosine-bornite intergrowth are figured in plate XXI, figs. 5 and 6.

5. Paragenesis of the skarn zones.

The skarn zones have been divided broadly into 2 groups, each characterized by a distinctive set of minerals. Skarns of group I are found immediately in contact with the granite and typically carry no fluorine- or boron-bearing minerals. Those of group II occupy a zone

n the dolomite side of the skarns of group I and are the home, not only
the fluorine-boron-bearing assemblages, but of monticellite and of the
hief ore mineral magnetite.

The development of the skarns of group I may be unaccompanied by
ose of group II (loc. K, quarries near no. 2 Prospect, Kilchrist). They
e to be attributed to a local silication of the dolomite, the effects of
hich can often be related to endomorphic changes in the contact facies
the granite itself. Broadly we may include among skarns of group I
e successive zones arranged as follows:



s has already been described, the association of the grossular-wollas-
nite zone at Kilchrist and at loc. K on the south side of the granite is
ith an endomorphic facies of granite notably potassic and carrying an
kali-clinopyroxene. The hedenbergite-plagioclase skarn development
associated with a contact facies of granite enriched in plagioclase giving
lear evidence that lime has been incorporated in the granite (pl. XXI,
g. 4). At Kilbride (loc. E) the contact phase of the granite assumes
gainst a broad zone of hedenbergite-plagioclase skarn, locally the
haracter of a quartz-monzonite (table XII, no. 5).

Confining our attention to the metasomatic phenomena of the skarn
ones at the contact itself, the contrasting compositions of the alterna-
ive first skarn layers may be illustrated by the analysed zones of (1)
rossular-wollastonite at Kilchrist, (2) hedenbergite-plagioclase at Kil-
ride (table IX).

Assuming equality of volume of skarn zone and dolomite from which
t is formed it is seen that in both cases, besides the main addition of
ilica and alumina, there has been almost a complete elimination of
magnesia. In the second example, however, there has been notable
ntroduction of ferrous oxide. Some lime has been eliminated, but in
he first example the reciprocal changes have involved the introduction
f lime which must have come from the dolomite beyond. The magnesia
liminated presumably moved out into the dolomite in the first example
where there is evidence of one way transfer only from the granite
Tilley, 1949).

In the case of the hedenbergite-plagioclase skarn some of the elimin-
ted lime may have moved into the granite as this has suffered an
ndomorphic change involving the production of plagioclase. But it is

TABLE IX. Analyses of skarns of group I.

	1.	2.	3.	Gain or loss in grams per 100 c.c. 3 → 1.	3 → 2.
SiO_2	46·48	46·62		+147·34	+156·27
Al_2O_3... ...	8·04	7·83	—	+25·48	+26·25
Fe_2O_3... ...	0·92	0·25	—	+2·91	+0·83
FeO	0·36	19·60	—	+1·14	+65·70
MnO	0·19	0·81	—	+0·60	+2·71
MgO	0·39	0·81	21·74	−61·16	−59·68
CaO	43·24	21·58	30·43	+49·74	−15·00
Na_2O	nil	0·66	—	—	+2·22
K_2O	0·04	0·54	—	—	+1·81
H_2O- ...	0·02	0·03	—	—	—
H_2O+ ...	0·51	0·86	—	—	—
TiO_2	0·05	0·07	—	—	—
F	—	0·98	—	—	+3·28
CO_2	—	—	47·82	−137·24	−137·24
	100·24	100·64, less 04·1 O=F, 100·23			
Sp. gr. ...	3·170	3·352	2·87		

1. Grossular-wollastonite skarn (first zone against granite), no. 1 Prospect, Kilchrist.
2. Hedenbergite-plagioclase skarn (first zone against granite), loc. E, Kilbride.
3. Dolomite (theoretical).

not easy to assess the balance of the transfer of lime to the granite as it seems probable that the skarn development completed itself before the final consolidation of the granite and some of the skarn itself may have been bodily incorporated in the contact facies. Whether any part of the magnesia moved out into the dolomite as in the first type of skarn cannot be determined.

These first-zone skarns are succeeded by a broader zone of clinopyroxene very characteristically divisible into a narrower zone green in colour and finer grained, and a wider pale clinopyroxene zone coarser in grain-size (cf. fig. 11, p. 650). Examples of the pyroxenes of these two zones at Kilbride (loc. D) contain respectively Fe oxides 2·28, 1·75 %, and alumina 6·42, 3·67% (nos. 20 and 19, table XIV). In this zonal arrangement we see a gradual decrease in the amounts of iron and aluminium fixed as the solutions moved out into the dolomite from the granite. Apart from that feature these clinopyroxenes derive from the straight silication of dolomite. These zones are usually, but not invariably, followed by a zone of varying width composed of forsterite with which some xanthophyllite may be associated.

By analogy with the succession seen around 'sponge forms' in the main aureole we may attribute this zone to processes operating during the formation of the group I skarns, and this zone therefore forms a component of the group, and the external one. The formation of this forsterite zone involves the continued outward movement of silica into the zone, addition of magnesia, and removal of lime. This addition of magnesia must be regarded as a process not necessarily conditioned by the availability of magnesia released from the first zone we have already remarked, for the zone is, as we have seen, typical of the halo around the 'sponge forms' in the main aureole. Reference to this particular problem will be made in the following section where the parageneses of the skarns of group II are considered.

Skarns of group II including the ore skarns.

Here we include the skarn zones found on the dolomite side of those just discussed. Besides magnetite they carry the group of boron and fluorine minerals of the mineral table I (p. 634) summarizing the minerals of the skarns.

Perhaps the simplest development of these skarns is seen at Kilchrist where the clinopyroxene zones of group I are succeeded by an ore skarn built of magnetite, forsterite, chondrodite or clinohumite, and diopside, together with, in places, accessory grossular and secondary clinochlore (fig. 1 *b*). The analysis of such a magnetite-diopside-chondrodite skarn is given in table XIII, no. 8, the composition of the magnetite concentrate from it under no. 9 in the same table. The analysed skarn itself contains approximately magnetite 78, diopside 11, and chondrodite 8 %. The analysis of the magnetite fraction shows that the ore mineral contains MnO 4 % and holds also some magnesia.

The occurrence of manganese and magnesia in skarn magnetites has previously been recorded, for example in the Quebec occurrence analysed by Harrington (1907). The ore skarn gives way sharply to the normal metamorphosed dolomite with forsterite and calcite. The development of the ore skarn of Kilchrist must be attributed to ore-bearing solutions, not directly local in origin, probably ascending from below and attacking preferentially the dolomite which has acted as an absorption apparatus.

The more complex skarn development of group II type at Kilchrist (loc. B) calls for some consideration. Here a zone rich in monticellite intervenes between the clinopyroxene zone with its forsterite edge and

the typical ore-skarn zone (fig. 1 *c*). The zone is built of monticellite, diopside, forsterite, and accessory clinohumite. Magnetite is variable in amount. Both the occurrence of this mineral and clinohumite point to this zone as a group II assemblage. Monticellite frequently invests both diopside and forsterite and may form large poikiloblasts with these minerals as enclosures. An analysis of a typical member of the skarn zone is given in table XIII, no. 10. The mineralogical composition is computed as follows: monticellite 50, diopside 13, forsterite 15, clinohumite 4, magnetite 16, calcite 1 %. This zone is clearly derived by the silication of a dolomite with the introduction of iron and probable removal of some lime.[1] Compared to the assemblages in the main aureole the zone shows the development of the normally higher index mineral of the decarbonation series, monticellite, as represented in the stage (*b*) of the equation

$$\underset{(a)}{CaMgSi_2O_6} + Mg_2SiO_4 + 2CaCO_3 \rightleftharpoons \underset{(b)}{3CaMgSiO_4} + 2CO_2$$

This question is further discussed in the sequel under the heading—role of monticellite.

The monticellite-bearing zone is followed outward by a magnetite-forsterite zone (with accessory clinohumite). In this skarn locality there is very clear evidence that the ore zone bears an invasive replacement relation to the monticellite zone, for in the sections the ore skarn zone can be seen veining it (cf. fig.1 *c*) and the skarns shown in figs. 12 and 13.

The magnetite-forsterite zone is thus the latest member of the skarn group. Beyond this zone appears marble with clinohumite passing out into the normal forsterite-marble. A variant of the monticellite-type skarn zone of this locality appears in the skarn succession at Kilbride (loc. E) fig. 1 *d*. There an intermediate zone reaching up to $4\frac{1}{2}$ inches in width is built of chondrodite (locally clinohumite), diopside, monticellite (variable in amount), and grossular. Here again the monticellite may appear in large poikiloblasts investing chondrodite and diopside. Grossular appears both as large grains and as an investing skin to grains of the other silicates as if developed late in the succession. Sections of this zone are figured in pl. XIX, figs. 3 and 4, showing the relations of grossular, and in pl. XX, figs. 3, 4, and 6. This zone is followed outwards by the idioblastic forsterite zone carrying xanthophyllite, often in process of replacement by magnetite (pl. XVIII, fig. 5).

[1] After removal of magnetite, the composition plotted in the diagram of fig. 14 falls at B in the appropriate triangle.

In the neighbouring skarn section of loc. D *b*, Kilbride, the skarn zones are succeeded by an inconstant monticellite zone and a wide zone built essentially of diopside, cuspidine, and clinohumite; this is followed by a second monticellite zone, an inconstant clinohumite zone, and finally the dolomite. An analysis of the main diopside-cuspidine-clinohumite zone is given in table XIII, no. 13. Its computed composition is cuspidine 50, clinohumite 35, diopside 15 %. The ratio CaO:MgO = 1·12:1 is that of a dolomite and the ratios CaO:(Mg,Fe)O:SiO_2 have been plotted at A, in fig. 14, falling within the triangle cuspidine-diopside-clinohumite.

A somewhat similar zone occurs in the skarn succession at Camas Malag (loc. G *a*). There the ore-bearing solutions have used the contact surfaces of a dolerite dike in the marble as channel ways and converted the marble into a diopside-cuspidine-magnetite skarn reaching a thickness of two feet (fig. 4). This zone is followed outwards by a broad monticellite zone carrying magnetite.

The role of monticellite in the aureole.

Monticellite is unknown in the aureole beyond the immediate vicinity of the skarn zones. Monticellite-bearing marbles appear immediately adjacent to the monticellite zone of Camas Malag, and at Kilbride a similar assemblage is met with in like association, and as narrow brown stringers through a diopside-forsterite-marble (loc. E). These stringers form vein-like masses through the marble investing forsterite grains or cutting diopside hornfels (pl. XVIII, fig. 6). It may be accompanied by accessory cuspidine.

That the development of monticellite in this environment is late and due to solutions connected with the formation of the skarns is clear from these relationships. In the aureole proper the highest grade of metamorphism is revealed in the development of periclase and in the rare presence of wollastonite.

The widespread development of monticellite in the skarn zones is, however, not necessarily to be attributed to higher temperatures, for in the skarn environment metamorphism and metasomatism is accomplished by moving solutions and the entry of this normally higher stage in the decarbonation series may quite reasonably be attributed to reduction of pressure of CO_2 due to its escape in such solutions.

There remains some discussion of the time relations of the development of the monticellite-rich zones to those by which they are bounded. At Kilchrist (loc. B) the monticellite-diopside-forsterite zone was probably

formed at an early stage among the assemblages of group II skarns. It is followed outward by the forsterite-magnetite zone carrying in places clinohumite. This zone shows a distinct transgressive relation to the monticellite zone and has in part been formed by metasomatism of this zone (figs. 12 and 13).

At Kilbride (loc. D *b*) there are two prominent monticellite zones, the inner adjacent to a narrow clinohumite zone (fig. 2). This fluosilicate

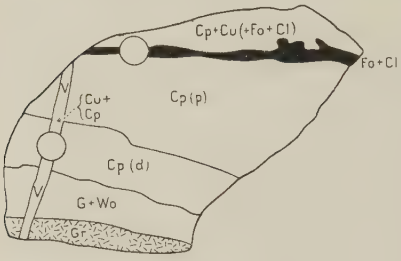


Fig. 11.

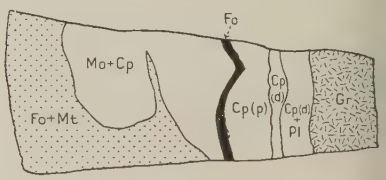


Fig. 12.

Fig. 11. Zoned skarn at granite contact, Kilbride (loc. D *b*). ×½.

Clinopyroxene granite (Gr), grossular-wollastonite zone (G + Wo), dark and pale clinopyroxene zones (Cp(d) and Cp(p)), forsterite-clinohumite zone (Fo + Cl).

A later vein cross-cutting the skarn zones consists in part of idocrase (V) and clinopyroxene (Cp), in part of cuspidine (Cu) and clinopyroxene.

The ringed areas are figured in pl. XVIII, fig. 3 (upper ring), fig. 4 (lower ring).

Fig. 12. Zoned skarn, loc. B, Kilchrist. ×½.

A zone built of hedenbergite and plagioclase (Cp(d) + Pl) borders the pyroxene-bearing granite; there follow two clinopyroxene zones, a forsterite zone, clinopyroxene-monticellite zone (Mo + Cp), and finally, a transgressive ore skarn built of magnetite and forsterite with scattered grains of bornite and chalcosine.

zone appears to have been formed by transformation of the forsterite zone which normally succeeds the two clinopyroxene zones of group I assemblages (pl. XVIII, fig. 3). The inner monticellite zone can be traced in a transgressive band through the broader diopside-cuspidine zone, appearing as a narrow monticellite band carrying cuspidine with a fine-grained monticellite edge corroding the large diopside crystals of the diopside-cuspidine zone (pl. XXI, fig. 1). On the marble side of this zone the second or outer monticellite zone is developed. This may make direct contact with the marble carrying clinohumite or chondrodite and spinel, but the section also reveals that a second clinohumite zone (½ inch thick) may intervene in places between the marble and the monticellite zone.

Study of the contact of these monticellite and clinohumite zones shows hat the monticellite grains typically invest grains of clinohumite and ay even form an oriented overgrowth on clinohumite crystals (pl. XX, g. 3). Again, where the monticellite zone is in contact with the marble, he monticellite encloses resi- ues of clinohumite grains and pinel, both present in the adja- ent marble. There is evi- nce, therefore, that part at ast of these monticellite zones s arisen subsequent to the velopment of the fluosilicate- aring zones.

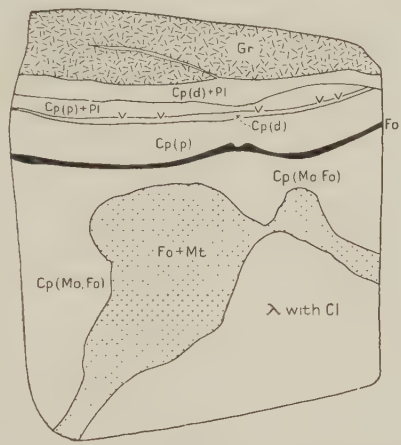


Fig. 13. Zoned skarn against clinopyroxene granite (Gr), loc. B, Kilchrist. $\times\frac{1}{2}$. Succession of zones as shown with a thin idocrase zone (V). The transgressive ore skarn, forsterite and magnetite (Fo+Mt) against the clinopyroxene-monticellite-forsterite assemblage Cp (Mo, Fo) is followed by marble with idioblastic clinohumite (λ with Cl).

At Camas Malag (loc. G *a*) he diopside-cuspidine zone is contact with a monticellite- agnetite skarn, and here, too, here is evidence that some of he monticellite development is ate in the sequence of meta- omatic changes. Monticellite, is clear, develops by trans- rmation of dolomite itself, ay be later than the develop- ent of the fluosilicates, and ven arise at the expense of arlier formed zones of the type f diopside-cuspidine-clinohumite. The analysis of that zone from Kil- ride shows that except for the fluorine content it matches closely a onticellite composition (table XIII, no. 13).

With the above reservation, monticellite on the one hand, and an ssemblage built largely of cuspidine and an olivine or humite mineral n the other, may be quasi-heteromorphic assemblages (fig. 14).

$$\begin{matrix}4CaMgSiO_4\\2HF\end{matrix}\Big\| Ca_4Si_2O_7F_2+2Mg_2SiO_4+H_2O.$$

he cuspidine-forsterite-(clinohumite) assemblages of Camas Malag loc. F) are conspicuous examples of such an association, heteromorphic f monticellite (pl. XIX, figs. 1 and 2). Though evidence for the direct ransformation of monticellite to a fluosilicate assemblage has been

sought, none has been found, and the lack of such evidence is consonant with a relative late development of monticellite.

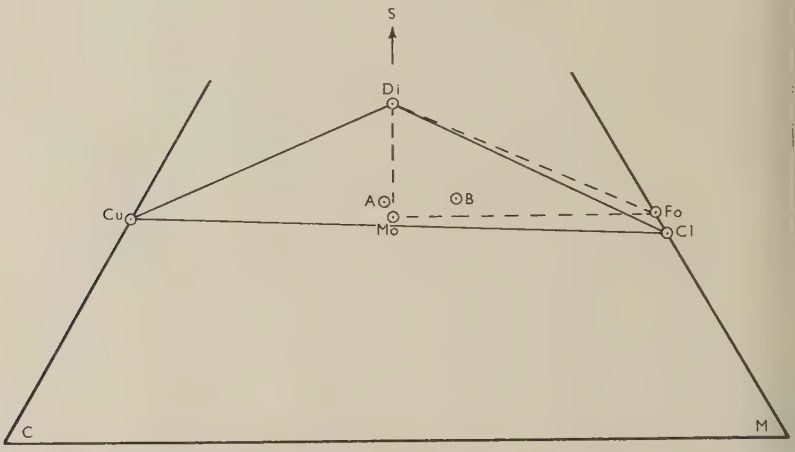


Fig. 14. Plot showing the compositions of the skarn zones diopside-cuspidine-clinohumite (A) and diopside-monticellite-forsterite (B) in the system CaO–(Mg,Fe)O–SiO_2 (CMS). Cu cuspidine, Cl clinohumite, Mo monticellite, Fo forsterite.

6. Products of lime metasomatism.

(*a*) *Metasomatism of pre-granitic dolerite dikes.*

Dolerite dikes invading the Durness dolomite horizons are cut off abruptly at the contact of the granite and are contact-altered. The petrography of these pre-granitic dolerites has been described by Harker in the Skye memoir (1904, pp. 318, 320), and he has discussed the changes they experience in the granite aureole. Further changes of a fundamental kind affect these rocks where they are associated with the ore skarns. Probably the best development of this metasomatism is to be found at no. 1 Prospect, Kilchrist, at Camas Malag, and at locality H on the south side of the granite.

The dolerites are ophitic to subophitic types, ordinarily containing no olivine. They consist of augite, labradorite, and subordinate iron ores. In the aureole they suffer amphibolization usually without their doleritic textures being effaced. In an example from no. 1 Prospect, Kilchrist, a 15-inch dike striking towards the granite contact (sketch-map, fig. 15), the augite is partially or wholly replaced by a brown pleochroic amphibole. It may become mottled with this new hornblende, be wholly replaced by it, or be reduced to a granoblastic aggregate consisting of interlocking grains of pyroxene and amphibole. The plagioclase remains

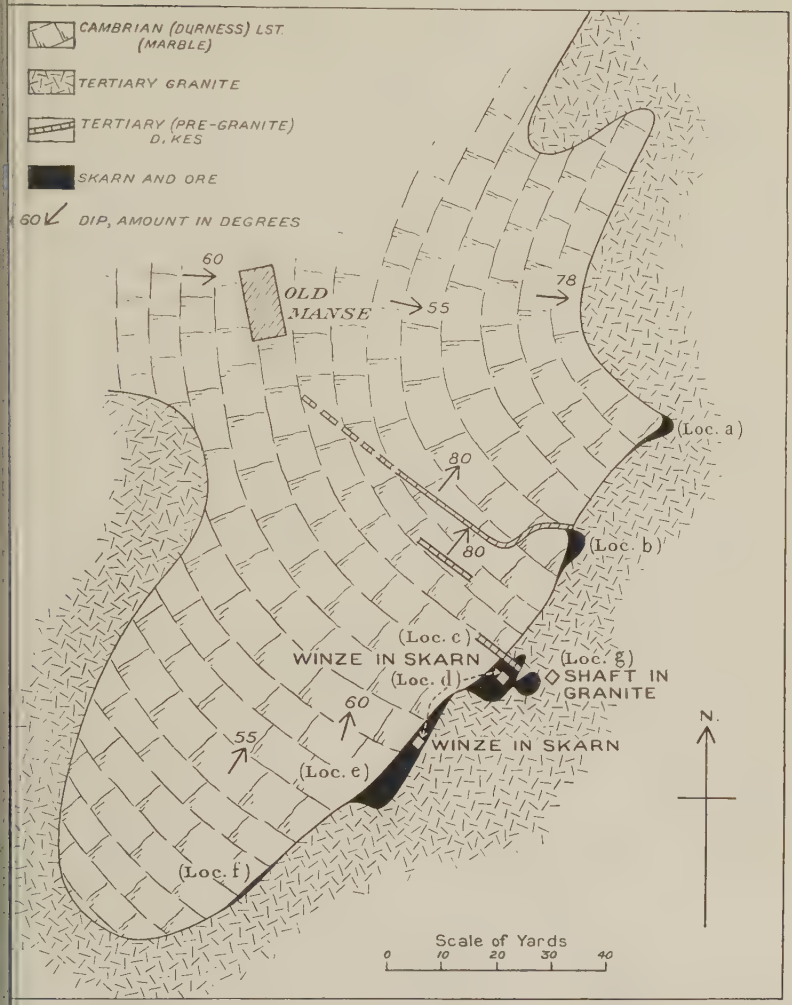


FIG. 15. Sketch-map of no. 1 Prospect (locality A, sub-localities *a–g*), Kilchrist, Skye.

unchanged, of composition $Ab_{40}An_{60}$. This amphibolized dolerite has been analysed (table X, no. 1). It may be compared with an analysis of an unaltered olivine-free dolerite of an inclined sheet intersecting the Cuillin gabbro (Harker, 1904, p. 370).

The radical change in chemical composition and mineralogy suffered by these dolerites when they become involved in the ore skarn environment is shown in a specimen of an adjacent dolerite at Kilchrist associated with the magnetite ore skarn ('*c*' section, loc. A). A specimen of this rock showing the effects of metasomatism in hand-specimens is described below. The specimen comes from a dike entangled in the ore skarn at the granite contact, and microscopic study shows that it has suffered strong metasomatism. It is crossed by narrow venules bearing diopside and idocrase, and against these venules the transition to amphibolized dolerite can be traced. The venule is built at its core of diopside (2 mm.) and has a margin of idocrase on both sides (1·5 mm.). In this part of the dolerite all original texture has been destroyed.

Beyond the idocrase bands metasomatism has not completely effaced original textures. The assemblage is then built of idocrase and diopside, the plagioclase laths being replaced pseudomorphously by idocrase of the same type as in the marginal vein. Traced outward this idocrase replacement is succeeded by that of grossular, the rock then being built of clinopyroxene aggregates with labradorite laths completely converted to garnet (pl. XXII, figs. 1–4). More remote from the venule brown hornblende is found among the clinopyroxene, and eventually labradorite laths remain unchanged or are partially replaced by sericite in aggregates of flakes. Biotite may be sporadically developed.

The transformation stages can thus be followed in their entirety from the amphibolized dolerite to the venule where all structures are lost and the rock consists of diopside and idocrase. Similar transformations can be followed from the ore skarn contacts. Elsewhere the dolerite may be completely replaced by diopside in microgranulitic aggregates ± grossular, and in some examples this aggregate may be crossed by thin venules of xanthophyllite carrying sporadic green spinel which may also appear dispersed in the clinopyroxene groundmass.

An analysis of the metasomatized dolerite in the ore skarn at '*c*' section Kilchrist (loc. A) is set out in table X, no. 2. The material analysed is a garnetized dolerite showing conspicuous laths of plagioclase in all stages of conversion to grossular, the ferromagnesium minerals being both augite and hornblende. Some relatively unaltered areas in which labradorite is partially replaced by sericite are also included in the material analysed. The analysis is to be compared with that of the adjacent 15-inch dike described (table X, no. 1). The preservation of the doleritic texture in the metasomatism allows us to set out the changes

volved, by comparing equal volumes of the two rocks, the gains and sses being shown in column 3 (table X).

TABLE X. Analyses of dolerite and dolerite skarn, Kilchrist.

	1.	2.	3. Gain or loss in grams per 100 c.c. 1 → 2.
SiO_2	45·90	43·86	+5·60
Al_2O_3	15·68	14·49	+0·29
Fe_2O_3	2·29	2·34	+0·78
FeO	7·89	6·39	−2·82
MnO	0·12	0·37	+0·86
MgO	8·11	7·41	−0·13
CaO	12·61	18·75	+23·62
Na_2O	2·96	1·03	−5·56
K_2O	0·45	1·25	+2·76
TiO_2	1·89	1·74	+0·01
CO_2	0·58	0·23	−1·00
H_2O-	0·25	0·19	
H_2O+	0·85	1·49	
S	0·92	0·71	
	100·50	100·25	
Less O=S... ...	0·46	0·35	
	100·04	99·90	
Sp. gr.	3·026	3·295	

1. Amphibolized dolerite, 15-inch dike close to granite, no. 1 Prospect, Kilchrist (loc. A).
2. Dolerite skarn (garnetized dolerite hornfels), associated with ore skarns, no. 1 rospect, Kilchrist.

The principal changes there noted are the introduction of lime, silica, nd potash, and removal of soda and ferrous oxide. These changes find dequate expression in the grossular development from labradorite, he presence of subordinate biotite, and the sericite replacing plagioclase.

(*b*) *Metasomatism of the granite.*

Endomorphic skarn formation in the granite is found at the contact t no. 1 Prospect, Kilchrist. It is local in its development, being asso- ated with late vein formation in the vicinity of the ore skarns. xamples are recognized from two points at Kilchrist, one close to the ' section (fig. 15) where it occurs close to the granite-skarn junction at he ledge and in specimens from the trial pit in the granite close to '*c*' ection.

The granite is traversed by dark zones or veins which prove to be nes along which solutions have penetrated and caused fundamental

changes in the mineralogy (fig. 16). The width of these bands varies, ranging from five inches down to a fraction of a millimetre. In hand-specimens the metasomatic band appears often sharply, if irregularly, separated from the lighter coloured granite.

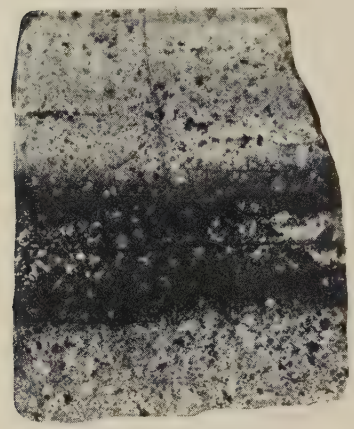


Fig. 16. Metasomatic vein in granite, no. 1 Prospect, Kilchrist. $\times\frac{2}{3}$.

Hornblende-biotite-granite rich in potash-felspar veined by solutions giving rise to a clinopyroxene-oligoclase metasomatic product with accessory fluorite and with areas rich in grossular-andradite and fibrous idocrase. The granite and vein product have been analysed and figured.

The constituent minerals of the veins are clinopyroxene (diopside to hedenbergite), plagioclase (oligoclase to andesine), fluorite, grossular-andradite, idocrase, and accessorily epidote and orthite. In the narrowest veins the simplest expression of the metasomatism is the appearance of a narrow band of green clinopyroxene (near hedenbergite) with which a little fluorite is frequently associated. The commonest skarn assemblage is, however, in the form of medium-grained aggregates of clinopyroxene with twinned andesine or oligoclase, again usually with some fluorite. These are seen to end sharply against the granite. More complex veins carry grossular-andradite and idocrase, while locally orthite may be prominent.

Analyses have been carried out on the normal granite and an associated metasomatic vein. The granite (table XI, column 1) bears oligoclase, microperthite, quartz, biotite, and hornblende, also as accessories, magnetite, sphene, and grains of orthite. This is veined for a width of $1\frac{1}{2}$ inches. The onset of the metasomatized state is rather sudden. Potash-felspar and quartz disappear and a clinopyroxene (hedenbergite) is developed surrounding the plagioclase, oligoclase or oligoclase-andesine, often speckled with fine calcite.

The composition of the plagioclase is not uniform throughout the vein; in areas where garnet is developed a striated albite may appear. Idocrase is abundantly developed in some sections in the form of a mat of light green, optically negative, fibres which appear to be developing at the expense of the felspar. Orthite is sporadic, in part intergrown with yellow epidote, often strongly zoned and otherwise patchy.

This orthite is an optically negative type with dispersion $r < v$, pleochroism variable, but β and γ are brown, while α is yellow or brownish-green. Absorption $\gamma > \beta \geqslant \alpha$.

In the final stages of metasomatism the centre of the vein is composed of an idocrase-clinopyroxene skarn where idocrase has completely replaced the plagioclase. The successive stages in the replacement process are depicted in the plate of photomicrographs (pl. XXIII, figs. 1–6).

The analysis of the vein material, essentially clinopyroxene, andesine, fluorite, some grossular, and a little idocrase, is set down in column 2 (Table XI). Assuming no change of volume in the transformation we see

TABLE XI. Analyses of granite and associated metasomatic vein, no. 1 Prospect, Kilchrist, Skye.

	1.	2.	3. Gain or loss in grams per 100 c.c.
SiO_2	74·88	47·65	−57·66
Al_2O_3	12·73	11·02	−1·29
Fe_2O_3...	0·53	5·88	+15·75
FeO	1·33	6·72	+16·10
MnO	0·02	0·85	+2·40
MgO	0·25	0·29	+0·19
CaO	1·12	19·33	+53·40
Na_2O...	3·33	2·44	−1·63
K_2O	4·99	0·87	−10·57
TiO_2	0·18	0·09	
P_2O_5	0·05	—	
H_2O-	0·16	0·38	
H_2O+	0·79	1·30	
F	—	0·95	+2·77
CO_2	—	2·69	+7·84
	100·36	100·46, less O for F 100·06	
Sp. gr.	2·625	2·915	
Norm:			
Quartz	32·94	6·42	
Orthoclase	29·47	5·56	
Albite	28·30	20·44	
Anorthite	5·00	16·40	
Diopside	0·71	18·36	
Hypersthene... ...	1·82	Wo 14·50	
Magnetite	0·70	8·58	
Ilmenite	0·46	—	
Fluorite	—	1·95	
Calcite	—	6·11	

1. Biotite-hornblende-granite.
2. Clinopyroxene-plagioclase vein with fluorite, grossular, and little idocrase.

that the significant chemical changes are removal of silica and potash and the accession of ferric and ferrous iron, and lime. Mineralogically we observe that quartz and potash-felspar disappear, potash being removed, but the alumina of the orthoclase building up with lime new plagioclase, while the lime and ferrous iron give the new-formed hedenbergite. The introduced ferric oxide appears principally in grossular-andradite.

The analysis of column 2 represents an intermediate stage of transformation. The clinopyroxene-idocrase assemblage—equivalent to the mineralogical combination ultimately found too in the skarned dolerite—denotes of course the final stage of lime metasomatism with the building up of idocrase (and garnet) at the expense of the plagioclase. Good examples of replacement of plagioclase by idocrase are figured in pl. XXIII, figs. 5 and 6.

7. Contrasts of lime and iron-magnesia metasomatism in the skarn successions.

The time relations of the primary (group I) and secondary (group II) skarn assemblages are manifest from their relationships. The first group may be unaccompanied by ore skarns, and the ore skarns themselves are not necessarily restricted to the environment of the first group which are spatially bound to the granite. Furthermore, the ore skarns are seen to traverse not only parts of their own succession, but also, though less commonly, the primary group I zones, as in the venules of magnetite skarn which cross-cut the diopside zones of this group. It seems evident that we may conclude that the skarns of group II result from solutions which have moved into their present position mainly through the dolomite bounding the primary skarns. The iron-bearing solutions found the carbonate-rocks more easily replaceable than the silicated group I assemblages.

In the primary zones the inner grossular-wollastonite assemblage represents a silicated dolomite to which alumina has been contributed from the granite and to which lime also has been added. It provides an example of lime metasomatism, even if only on a small scale. Lime metasomatism is prominent in the inner zones of 'sponge forms' in the skarns of group II. In nature they contrast with those of the 'sponge forms' in the main aureole. The latter are tremolitic and diopsidic, but in the former calcium minerals, wollastonite and cuspidine, are prominent, these eventually giving place externally to diopside. Lime metaso-

natism of the chert material or of the dolomite has, in the case of uspidine, been accompanied by introduction of fluorine.

These zones differ from those of the granite contact by the absence or arity of grossular. The silica of the cherts has contributed essentially to he inner zones around them, whereas with the granite, alumina as well as silica has been available.

The zones of the 'sponge forms' beyond have, however, clearly derived their silica from the solutions which are responsible for the ore skarns. It seems probable that the original size of these 'sponge forms' includes not only the quartz core which itself may be wholly replaced, but also the zones extending out to the limits of the diopside zone.

In the skarns of the second group, those rich in humite minerals, forsterite, and magnetite, the striking metasomatism is associated with the introduction of iron and magnesia and a notable loss in lime. The iron solutions are doubtless derived from residual solutions from the granite, but the source of magnesia may most likely be sought in the dolomite country-rock, ultimately to be precipitated, along with or separately from magnetite, in the humite minerals and forsterite.

The evidence for lime removal in the development of these ore skarns being thus clear, it may be inquired if there is evidence of the fate of these lime-bearing solutions. At Kilchrist it is apparent that these released solutions have themselves been responsible for late metasomatism.

As already described, the granite itself at the contact and the dolerite dikes entangled in the ore skarns have suffered locally a thoroughgoing lime replacement, and we may confidently attribute this process to the lime solutions set free in the ore skarn development. It has clearly operated subsequent to the consolidation of the border zone of the granite.[1]

8. Comparison with the skarn zones of other regions.

Examples of boron-fluorine metasomatism associated with banded skarns have not frequently been described. Perhaps the closest analogue to the Skye skarn development is that described by Watanabe (1943) from the Suian district of Korea as exemplified in the Hol Kol area, where dolomites have in places been converted at the contacts of the

[1] Umpleby (1914, 1917) has described the garnetization of granite-porphyry of the Mackay region, Idaho, around included blocks of limestone, and he associates the lime metasomatism of the granite-porphyry with the release of lime during the garnetization of the adjacent limestone inclusions (1914, p. 350; 1917, p. 62).

Suian granite into banded skarns bearing chondrodite minerals, and the boron minerals kotoite, fluoborite, and ludwigite.

In the Norberg district of Sweden, Geijer (1927, 1939) has described a somewhat similar boron-fluorine metasomatism with the production of humite minerals, ludwigite, and fluoborite in association with magnetite ore skarns in dolomite.

Apart from the occurrence of boron minerals, the Skye skarns in their multizoned character show resemblance to the ore skarns of Pitkäranta described by Trüstedt (1907) and to the more recently described banded skarns ('ribbon rock') of Iron Mountain, New Mexico (Jahns, 1944, *a*, *b*). While Trüstedt expressed the opinion that some of his skarns were due to replacement, he conceived for some of them deposition in open channels (by crustification). Such a mode of occurrence would require solutions of repeatedly changing composition, and the evidence that Trüstedt supplies for the operation of such a process is not entirely convincing. At Iron Mountain, New Mexico, Jahns has set out the evidence for a replacement origin of his rhythmically layered 'ribbon rock', and he remarks that this layering is probably best interpreted as a diffusion effect as in the Liesegang phenomenon. Moreover, the 'ribbon rock' occurs on the limestone side of the contact zone, i.e. between the limestone and the massive tactite which is nearest the intrusion. These conditions, as we have attempted to show, are reproduced in the development of the Broadford skarns.

Acknowledgements.—The map of plate XXIV is reproduced, with a few minor emendations, from the six-inch Geological Survey sheet (one-inch 71, six-inch 46, Scotland) mapped by Dr. A. Harker and Mr. C. T. Clough. For permission to make this reproduction the author is indebted to the Director of the Geological Survey.

Professor W. Q. Kennedy has kindly supplied the sketch-map of figure 15, and thanks are especially due to him for the opportunity of co-operating in the study of those skarn localities first discovered by him and for the fruitful discussion of genetic problems in the early stages of the work.

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TABLE XII. Analyses of the Beinn an Dubhaich granite and its modifications.

	1.	2.	3.	4.	5.	6.
SiO_2	74·88	73·99	75·33	74·69	70·72	47·65
Al_2O_3	12·73	13·02	11·88	12·00	12·64	11·02
Fe_2O_3	0·53	0·76	0·43	0·44	0·21	5·88
FeO	1·33	1·60	0·95	0·71	2·79	6·72
MnO	0·02	0·02	0·01	0·03	0·06	0·85
MgO	0·25	0·31	0·08	0·12	1·10	0·29
CaO	1·12	1·22	0·92	1·66	4·18	19·33
Na_2O	3·33	3·43	2·55	2·02	3·24	2·44
K_2O	4·99	4·74	7·18	7·78	3·40	0·87
TiO_2	0·18	0·32	0·06	0·14	0·36	0·09
P_2O_5	0·05	0·05	0·08	trace	0·10	—
H_2O-	0·16	0·14	0·06	0·20	0·22	0·38
H_2O+	0·79	0·22	0·23	0·17	0·55	1·30
CO_2	—	—	—	nil	0·44	2·69
... ...	—	—	—	—	—	0·95
	100·36	99·82	99·76	99·96	100·01	100·46
					less O for F	100·06

1. Biotite-hornblende-granite, no. 1 Prospect, Kilchrist (loc. A). (Tilley, 1949, p. 84).
2. Biotite-hornblende-granite, trial pit, no. 1 Prospect, Kilchrist (loc. A). (Tilley, 1949, p. 84.)
3. Aegirine-hedenbergite-granite, at contact with skarn, south side of Beinn an Dubhaich granite mass, ESE. of Camas Malag, Loch Slapin (loc. K). (Tilley, 1949, p. 84.)
4. Potash-rich granite, contact with skarn, no. 1 Prospect, Kilchrist (loc. A). (Tilley, 1949, p. 92.)
5. Clinopyroxene-hornblende-granite, bulk composition of band immediately adjacent to hedenbergite-plagioclase skarn, Kilbride (loc. E). Anal. J. H. Scoon.
6. Metasomatic vein (with clinopyroxene, plagioclase, idocrase, &c.) in granite (no. 1 of this table), no. 1 Prospect, Kilchrist (loc. A). Anal. Geochemical Laboratories.

TABLE XIII. Analyses of skarns of the Beinn an Dubhaich granite contact zone.

	7.	8.	9.	10.	11.	12.	13.	14.	15.	16.	17.	18.
SiO_2	46·48	8·96	1·76	33·23	46·62	30·53	37·53	5·47	38·48	12·94	43·86	45·90
Al_2O_3	8·04	3·61	2·64	1·16	7·83	2·63	1·55	2·62	1·99	2·61	14·49	15·68
Fe_2O_3	0·92	52·27	64·07	11·07	0·25	0·90	1·58	33·29	0·44	0·89	2·34	2·29
FeO	0·36	17·54	21·39	7·35	19·60	1·40	1·49	13·78	5·78	0·65	6·39	7·89
MnO	0·19	3·19	3·83	0·79	0·81	0·58	0·39	0·30	0·15	0·07	0·37	0·12
MgO	0·39	8·22	4·74	23·35	0·81	46·01	20·79	30·20	9·77	11·48	7·41	8·11
CaO	43·24	4·02	1·42	21·40	21·58	7·55	32·54	0·36	35·59	41·15	18·75	12·61
Na_2O	nil	—	—	—	0·66	—	nil	—	0·05	—	1·03	2·96
K_2O	0·04	—	—	...	0·54	—	nil	—	0·02	—	1·25	0·45
TiO_2	0·05	0·10	0·18	0·31	0·07	0·11	0·09	0·11	0·08	0·06	1·74	1·89
P_2O_5	—	0·07	—	—	—	—	—	0·01	0·05	0·06	—	—
H_2O-	0·02	0·05	—	0·16	0·03	0·14	0·15	0·09	0·16	2·64	0·19	0·25
H_2O+	0·51	1·33	—	1·10	0·86	3·89	1·19	2·47	nil	3·23	1·49	0·85
F	—	0·10	—	0·17	0·98	2·92	4·58	0·34	11·52	0·10	—	—
CO_2	—	0·85	—	0·44	—	4·37	0·07	—	—	16·48	0·23	0·58
B_2O_3	—	—	—	—	—	—	—	11·53	—	6·08	—	—
SO_3	—	—	—	—	—	—	—	—	0·38	—	—	—
S	—	—	—	—	—	—	—	—	—	0·55	0·71	0·92
	100·24	100·31	100·03	100·53	100·64	101·03	101·95	100·57	104·46	100·30	100·25	100·50
Less O = F, S ...	—	—	—	0·07	0·41	1·23	1·93	0·14	4·85	0·32	0·35	0·46
				100·46	100·23	99·80	100·02	100·43	99·61	99·98	99·90	100·04

7. Grossular-wollastonite skarn, no. 1 Prospect, Kilchrist (loc. A). (Tilley 1949, p. 88.)
8. Magnetite-chondrodite skarn with diopside, no. 1 Prospect, Kilchrist (loc. A). Anal. J. H. Scoon.
9. Magnetite concentrate from (8), same locality. Anal. J. H. Scoon.
10. Monticellite-diopside-forsterite skarn with magnetite and some clinohumite, contact of island of dolomite in granite, 300 yards ESE. of old Manse, Kilchrist (loc. B). Anal. J. H. Scoon.
11. Hedenbergite-plagioclase skarn, granite contact, Kilbride (loc. E). Anal. H. C. G. Vincent.
12. Clinohumite skarn (with some calcite, chlorite, xanthophyllite, and fluorite), Kilbride (loc. D *b*). Anal. Geochemical Laboratories.
13. Diopside-cuspidine-clinohumite skarn, Kilbride (loc. D *b*). Anal. J. H. Scoon.
14. Ludwigite-chondrodite skarn with magnetite, Kilbride (loc. D *a*). Anal. Geochemical Laboratories.
15. Clinopyroxene-fluorite skarn, Camas Malag (loc. F). Anal. Geochemical Laboratories.
16. Harkerite skarn, Camas Malag (loc. G *b*), with 1·31 CuO. Anal. Geochemical Laboratories.
17. Dolerite skarn (garnetized dolerite-hornfels) associated with ore skarns, no. 1 Prospect, Kilchrist (loc. A). Anal. J. H. Scoon.
18. Amphibolized dolerite, 15-inch dike, contact zone, no. 1 Prospect, Kilchrist (loc. A). Anal. J. H. Scoon.

TABLE XIV. Analyses of minerals from contact zones of the Beinn an Dubhaich granite.

	19.	20.	21.	22.	23.	24.	25.	26.
SiO_2 ...	50·74	49·81	49·69	36·55	50·00	31·31	1·28	14·17
Al_2O_3 ...	3·67	6·42	0·48	0·18	—	1·64	2·25	2·84
Fe_2O_3 ...	0·96	1·25	8·79	0·59	nil		35·31	0·85
FeO ...	0·79	1·03	18·23	4·71	9·29	—	16·36	0·46
MnO ...	0·08	0·06	0·25	—	1·22	—	0·29	0·02
MgO ...	16·92	15·22	1·98	22·11	nil	0·11	28·33	11·15
CaO ...	25·78	25·92	16·87	35·21	38·86	59·05	0·07	46·23
Na_2O ...	—	—	3·79	—	—	0·15	—	—
K_2O ...	—	—	nil	—	—		—	—
TiO_2 ...	0·11	0·14	0·45	nil	trace	—	0·30	—
P_2O_5 ...	—	—	—	—	—	—	—	—
H_2O- ...	0·11	0·06	nil	0·56	—	0·33	0·19	0·11
H_2O+ ...	0·54	0·13		0·84	—	0·41	0·44	0·81
F ...	—	—	—	—	—	9·80	—	—
CO_2 ...	0·24	0·15	—	nil	—	0·75	—	14·94
S_2O_3 ...	—	-	—	—	—	—	15·74	7·77
Cl ...	—	—	—	—	—	0·10	—	1·36
Insol. ...	—	—	—	—	0·45	—	-	—
	99·94	100·19	100·53	100·75	99·82	103·65	100·56	100·71
Less O = F, Cl	—	—	—	—	—	4·13	—	0·31
						99·52		100·40

19. Pale green clinopyroxene of zoned skarn, Kilbride (loc. D *b*); pale clinopyroxene zone. Anal. J. H. Scoon.
20. Green clinopyroxene of zoned skarn, Kilbride (loc. D *b*); dark clinopyroxene zone. Anal. J. H. Scoon.
21. Aegirine-hedenbergite, pyroxene of alkali-granite, ESE. of Camas Malag (loc. K); in granite (analysis 3 of table XII). (Tilley, 1949, p. 85.)
22. Monticellite of monticellite marble, Camas Malag (loc. G). Anal. Geochemical Laboratories.
23. Iron-wollastonite, in contact skarn, Camas Malag (loc. G). (Tilley, 1948 (*c*), p. 737.)
24. Cuspidine in cuspidine-clinopyroxene skarn, Camas Malag (loc. G). (Tilley, 1947, p. 93.)
25. Ludwigite in chondrodite-ludwigite-magnetite skarn, Kilbride (loc. D *a*). Anal. Geochemical Laboratories.
26. Harkerite in harkerite skarn with diopside, Camas Malag (loc. G). Anal. H. C. G. Vincent.

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Explanation of Plates XVIII–XXIV.

Plate XVIII. Kilbride marbles and skarns.

Fig. 1. Marble close to the granite contact, showing periclase with brucite haloes, spinel, and rounded grains of forsterite. (Loc. D *b*.) ×22. (Harker slide collection, 51881.)

Fig. 2. Marble with ludwigite (left centre), spinel, periclase with brownish haloes of brucite, rounded forsterite grains largely replaced by sheaves of clinohumite, and idioblastic clinohumite. (Loc. D *b*.) ×22. (51882.)

Fig. 3. Zoned skarn at contact. A zone of forsterite with intergrown clinohumite separates a diopside zone (right) from a broad diopside-cuspidine zone (left). Clear areas on the right edge of the central zone are secondary chlorite. (Loc. D *b*.) ×22. (52343/3.)

Fig. 4. Diopside-cuspidine vein cross-cutting the fine diopside zone of a zoned skarn. The positions of sections of figs. 3 4 are indicated by circles in the drawing of the zoned skarn of text-fig. 11. (Loc. D *b*.) ×22. (52343/4.)

Fig. 5. Idioblastic forsterite zone of zoned skarn. Closely packed forsterite crystals with a little interstitial calcite. Kilbride (Loc. E). ×17. (55768/1.)

Fig. 6. Monticellite border to diopside-rock at marble edge of zoned skarn. Fine diopside below, above coarser monticellite. (Loc. D *b*.) ×17. (55809.)

Plate XIX. Camas Malag and Kilbride skarns.

Fig. 1. Cuspidine-clinohumite-forsterite-calcite skarn. Spear-shaped crystals of cuspidine set in calcite and associated with clinohumite and forsterite. Camas Malag (loc. F). ×17. (55889/5.)

Fig. 2. Cuspidine-clinohumite-forsterite-calcite skarn. Another section of the same skarn: centre, large vertical section of cuspidine showing repeated twinning, surrounded by calcite (also twinned). The rest of the section shows clear cuspidine and grains of both clinohumite and forsterite. Camas Malag (loc. F). +nicols. ×17. (55889/4.)

Fig. 3. Chondrodite-grossular skarn. Idioblastic chondrodite wrapped round by grossular. Diopside is seen left of centre: some calcite. Kilbride (loc. E). ×17 (55748.)

Fig. 4. Chondrodite-grossular skarn. Shows the interstitial development of grossular (dark) in chondrodite-rock. Kilbride (loc. E). ×17. (55780/3.)

Fig. 5. Ludwigite-fluoborite skarn with magnetite. The clear areas are formed of fluoborite; dark, rich in acicular ludwigite; below some turbid areas of carbonate replacing chondrodite. Kilbride (loc. D *a*). ·17. (52303.)

Fig. 6. Ludwigite-chondrodite skarn with magnetite. Clear areas, chondrodite; dark, mostly acicular ludwigite with some intergrown magnetite. Kilbride (loc. D *a*). ×17. (52322/2.)

Plate XX. Kilchrist and Kilbride skarns.

Fig. 1. Clinohumite-diopside-forsterite-magnetite skarn. The section shows large clinohumite crystals enwrapped by magnetite. The rectangular section in the centre has a peripheral zone of more highly refringent forsterite in parallel growth. Kilchrist, no. 1 Prospect (loc. A). ×25. (50101.)

Fig. 2. Another section of the same skarn showing clinohumite cores to a forsterite grain. Both minerals are cut perpendicular to the acute bisectrix, one set of clinohumite lamellae being in the extinction position. Traces of the (010) cleavage of olivine at right angles to the twin lamellae of clinohumite are indicated. +nicols, ×66. (50104.)

Fig. 3. Outer monticellite zone of the banded skarn of Kilbride (loc. D *b*). The section shows a parallel orientated enclosure of clinohumite in a large crystal of monticellite. Clinohumite cut perpendicular to $\gamma(Bx_a)$, monticellite cut perpendicular to $\gamma(Bx_o)$. + nicols. ×70. (51869.)

Fig. 4. Twinned crystal of clinohumite in a clinohumite-monticellite-grossular skarn. The large crystal of clinohumite shown is cut perpendicular to the acute bisectrix and shows repeated twinning on (001) and a simple twin on (103) repeated in a set of fine lamellae on the right. Kilbride (loc. E), +nicols, +15. (55749/1.)

Fig. 5. Chondrodite-magnetite-skarn with diopside, Kilchrist (loc. A). + nicols, ×23. (49629.)

Fig. 6. Clinohumite-monticellite skarn. The upper half of the section represents a single crystal of monticellite at extinction, enclosing grains of clinohumite (poikiloblast): the lower half is made of clinohumite in diversely orientated grains. Kilbride (loc. E). +nicols, × 17. (55749/3.)

Plate XXI. Kilchrist, Kilbride, and Camas Malag skarns.

Fig. 1. Diopside-cuspidine-clinohumite skarn (right) bordered by a replacing skarn of monticellite with cuspidine (left). A fine edge of monticellite forms a sinuous border to coarse-grained clinopyroxene at the junction of the two skarns. Kilbride (loc. D *b*). ×17. (51883/3.)

Fig. 2. Harkerite-monticellite skarn. Idioblastic harkerite (with zones of secondary calcite), set in granular monticellite, opaque ore is an intergrowth of chalcosine and bornite. Camas Malag (loc. G *b*). ×22 (52190/2.)

Fig. 3. Hedenbergite-iron-wollastonite skarn. Showing the junction of a zone rich in iron-wollastonite solid solution (left) with a hedenbergite-granulitic plagioclase skarn. Kilbride (loc. E). ×17. (55752/2.)

Fig. 4. Contact of clinopyroxene-granite with hedenbergite skarn with granulitic plagioclase. Kilchrist (loc. B). ×17. (52354/3.)

Fig. 5. Monticellite-chalcosine-bornite skarn. Idioblastic monticellite set in copper ore, an intergrowth of chalcosine and bornite. Camas Malag (loc. G *b*). ×22. (50809.)

Fig. 6. A similar skarn, but here the ore enclosing monticellite consists of blende (centre) and chalcosine-bornite intergrowths. Camas Malag (loc. G *b*). ×22. (52181/1.)

Plate XXII. Amphibolized dolerite and dolerite skarn, no. 1 Prospect, Kilchrist.

Fig. 1. Amphibolized dolerite, 15-inch dike in marble. Clinopyroxene areas replaced by aggregates of brown hornblende. ×20. (52116.)

Fig. 2. Dolerite skarn entangled in ore skarn, 'c' section. Section from core of venule through dolerite, in succession from above: (1) clinopyroxene zone (Cp), (2) band of idocrase (V), (3) assemblage of grossular replacing plagioclase laths (G), and aggregates of turbid clinopyroxene. ×15. (52167.)

Fig. 3. The same. Below amphibolized dolerite passing upwards into dolerite skarn with plagioclase replaced by grossular (G) and hornblende aggregates by new clinopyroxene. ×15.

Fig. 4. The same. Dolerite completely converted to a skarn composed of grossular after plagioclase (G) and clinopyroxene. ×15.

Plate XXIII. Endomorphic skarns of Beinn an Dubhaich granite, no. 1 Prospect, Kilchrist (loc. A), Skye.

Fig. 1. Biotite-hornblende-granite, near metasomatic vein. The central dark grain is hornblende, associated with biotite, magnetite, and orthite; clear, quartz; turbid, perthitic orthoclase. Oligoclase appears below. ×17. (52134/8.)

Fig. 2. Clinopyroxene-plagioclase skarn, developed from rock of fig. 1 at the edge of a metasomatic vein. Aggregates of clinopyroxene with twinned andesine. Carbonate (calcite) is present in the clinopyroxene aggregates and epidote appears elsewhere in the section. ×17. (52134/3.

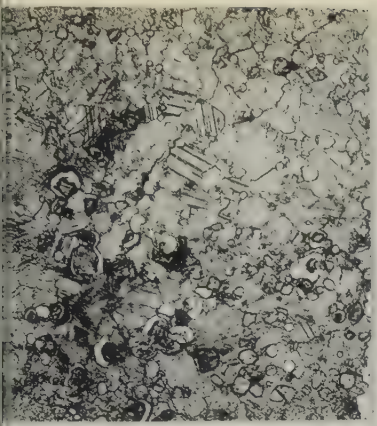
Fig. 3. Clinopyroxene-plagioclase skarn, developed from rock of fig. 1 in a metasomatic vein. In addition to pyroxene and andesine, clear fluorite is present. This rock has been analysed (table XII, no. 6). ×17. (52134/7.)

Fig. 4. Metasomatic vein in granite, showing zoned, twinned orthite associated with epidote and set in a matrix of clinopyroxene, oligoclase, and fluorite. Fibres of idocrase are developing in the plagioclase. ×23. (49678.)

Fig. 5. Idocrase-rich skarn. A further stage of metasomatism seen in the same rock as that of fig. 4. Fibres of idocrase replacing plagioclase are seen in the central portion of the photograph. ×23. (49684.)

Fig. 6. Idocrase-clinopyroxene skarn with orthite. A still further advanced stage, in which replacement of plagioclase by fibrous idocrase is complete. Centre and below orthite with some clinopyroxene. ×23. (49675.)

Plate XXIV. Geological and locality map of dolomite contact zones around the Beinn an Dubhaich granite, Skye.



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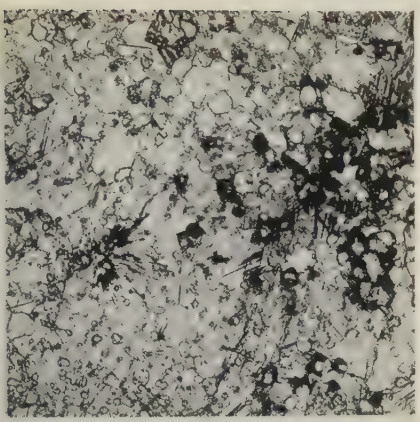

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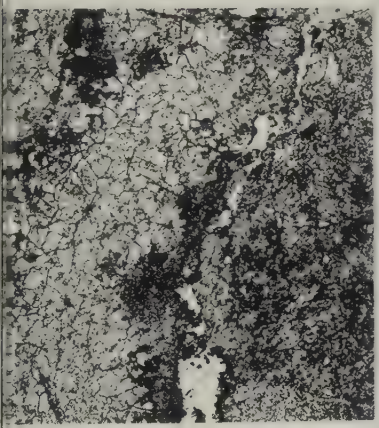

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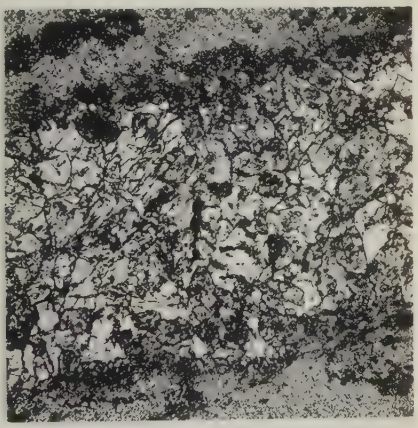

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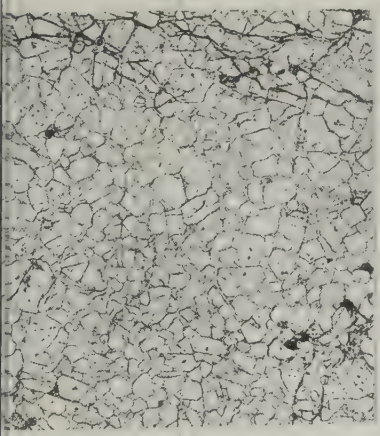

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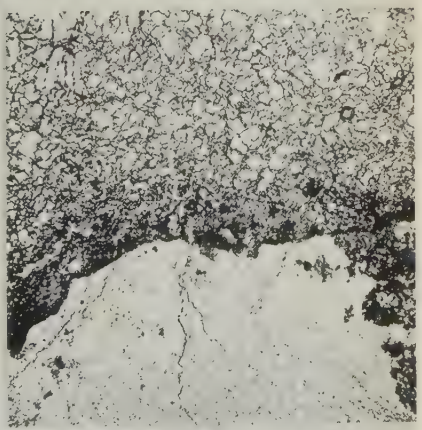

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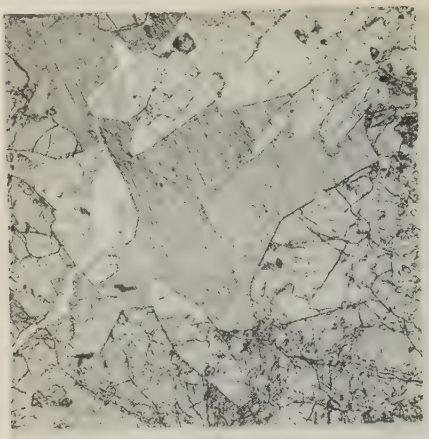
C. E. Tilley: Marbles and skarns, Kilbride, Skye



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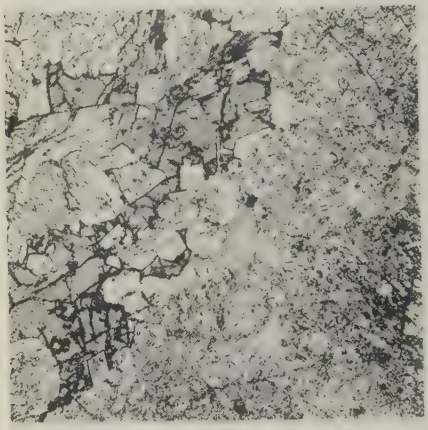

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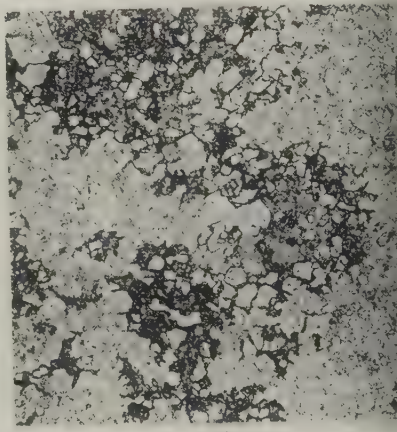

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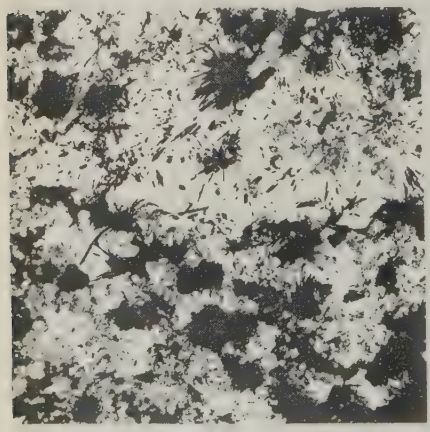

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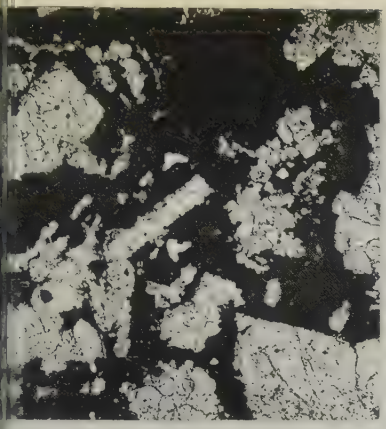
C. E. Tilley: Skarns, Camas Malag and Kilbride, Skye



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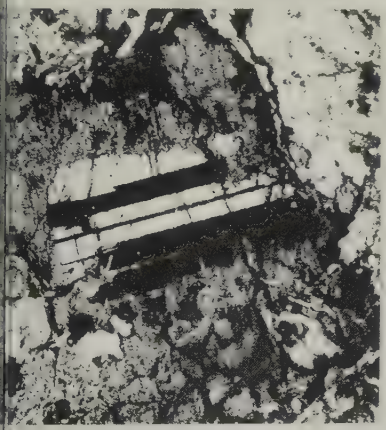

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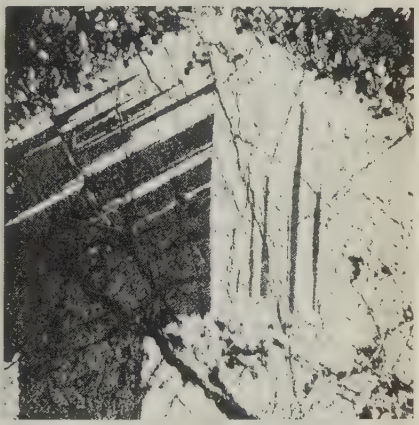

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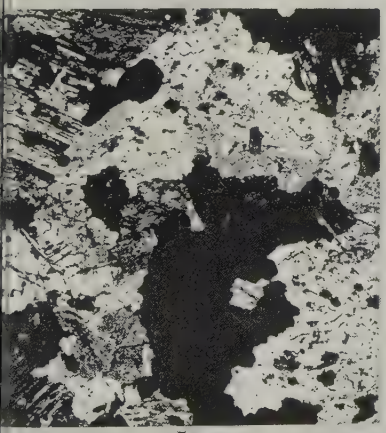

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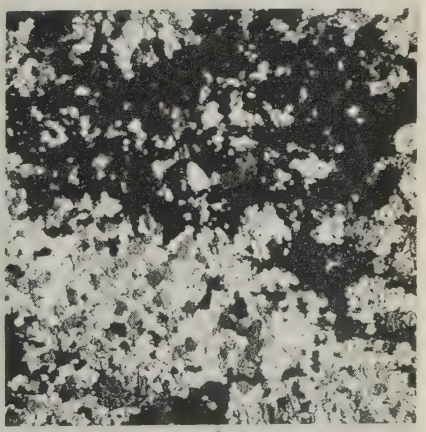

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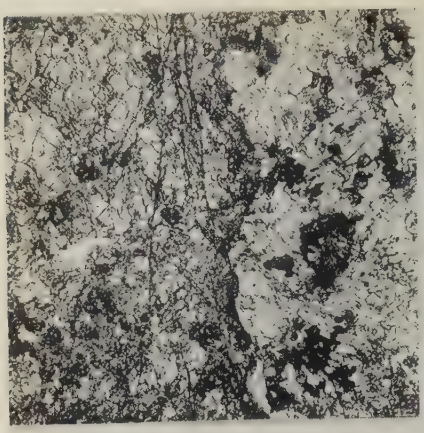
C. E. Tilley: Skarns, Kilchrist and Kilbride, Skye



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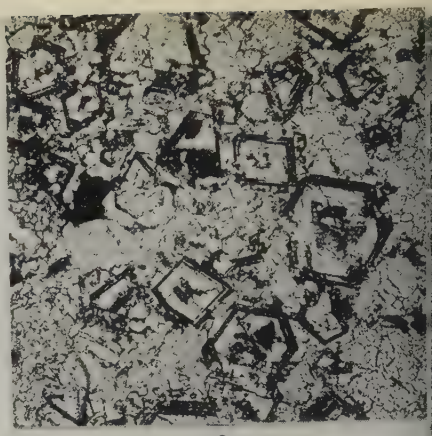

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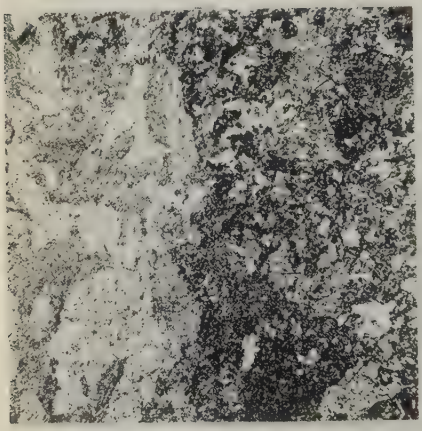

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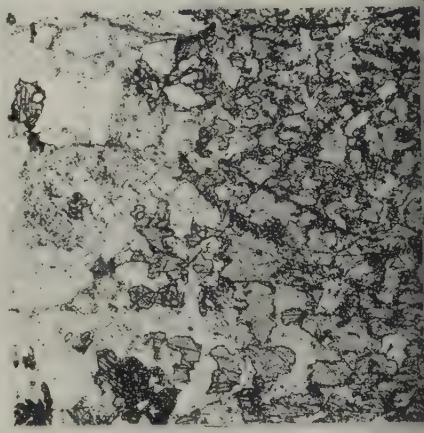

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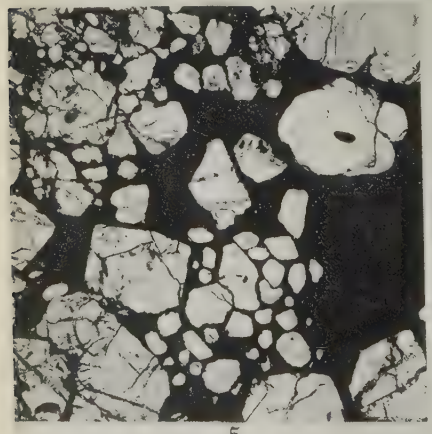

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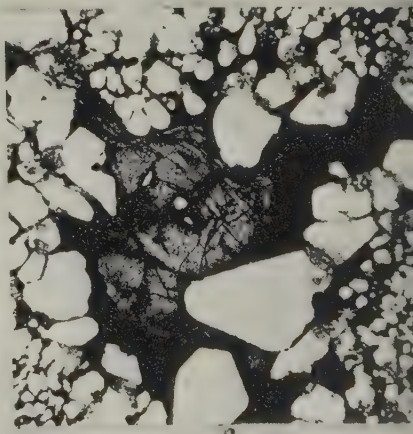

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C. E. Tilley: Skarns, Kilchrist, Kilbride, and Camas Malag, Skye

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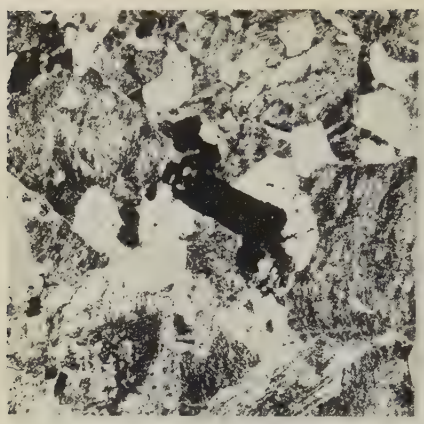
C. E. Tilley: Dolerite and dolerite skarn, Kilchrist, Skye



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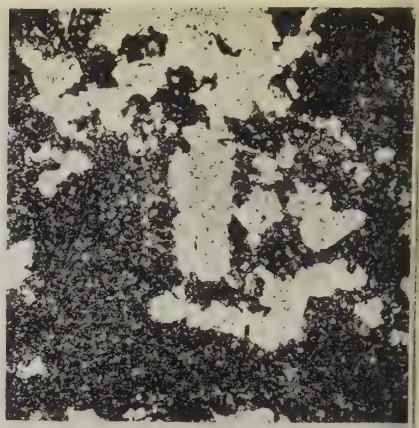

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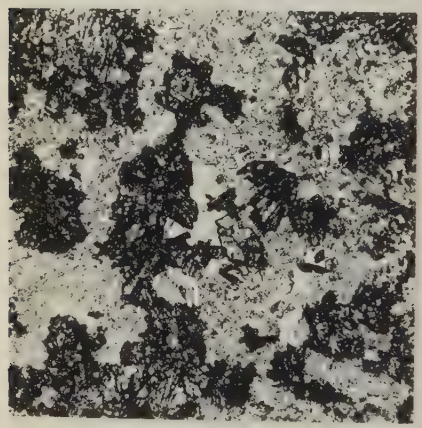

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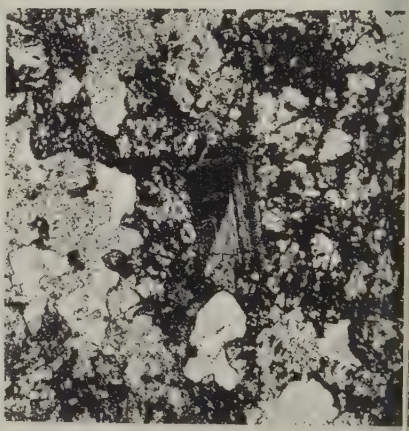

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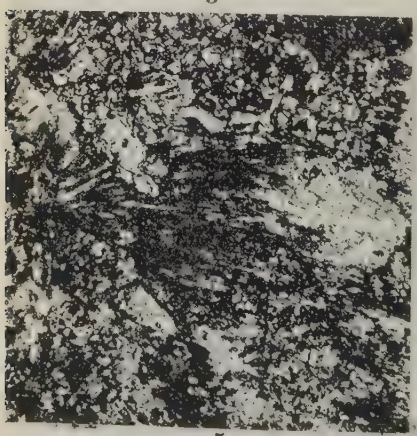

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C. E. TILLEY: ENDOMORPHIC SKARNS IN GRANITE, KILCHRIST, SKYE

C. E. TILLEY: GEOLOGICAL AND LOCALITY MAP OF DOLOMITE CONTACT ZONES AROUND THE BEINN AN DUBHAICH GRANITE, SKYE

The paragenesis of sylvine, carnallite, polyhalite, and kieserite in Eskdale borings nos. 3, 4, and 6, north-east Yorkshire.

(With Plate XXV.)

G. Armstrong, B.Sc., F.G.S., K. C. Dunham, D.Sc., S.D., F.G.S., F. O. Harvey, B.Sc., A.R.C.S., F.R.I.C., P. A. Sabine, Ph.D., A.R.C.S., F.G.S., and W. F. Waters, B.Sc., F.R.I.C.

Geological Survey of Great Britain.

[Read March 8, 1951.]

Introduction.

THE confirmation of the presence of potassium salts in the Permian evaporites beneath the Whitby district, first found in the D'Arcy Exploration Co.'s Eskdale no. 2 (Aislaby) boring (G. M. Lees and A. H. Taitt, 1945), and the discovery by Imperial Chemical Industries in their Eskdale no. 3 (Sleights), no. 4 (Sneaton), and no. 6 (Upgang) borings that these salts exist in workable quality and quantity, has been recorded in a recent historic paper by Dr. A. Fleck (1950). The successful outcome of this boring campaign may well lead to the outstanding development of the present century in economic mineralogy in Britain.

In the course of the exploration the Geological Survey has examined in detail the cores obtained, and every facility has been granted by Imperial Chemical Industries for the removal of specimens from the cores for preservation in the national collection at the Geological Museum. The Petrographical Department of the Geological Survey has, on its side, provided reports on the mineralogy of the cores. The present brief summary of these reports, with some additional chemical, X-ray, and petrographical data, is published by permission of the Director of the Geological Survey and with the approval of Imperial Chemical Industries, Limited.

Core-examination was carried out by Mr. G. Armstrong of the Newcastle Unit, successively under the direction of Dr. T. Robertson and Mr. W. Anderson; and by Prof. Dunham and Dr. Sabine of the Petrographical Department who were, respectively, responsible for the optical

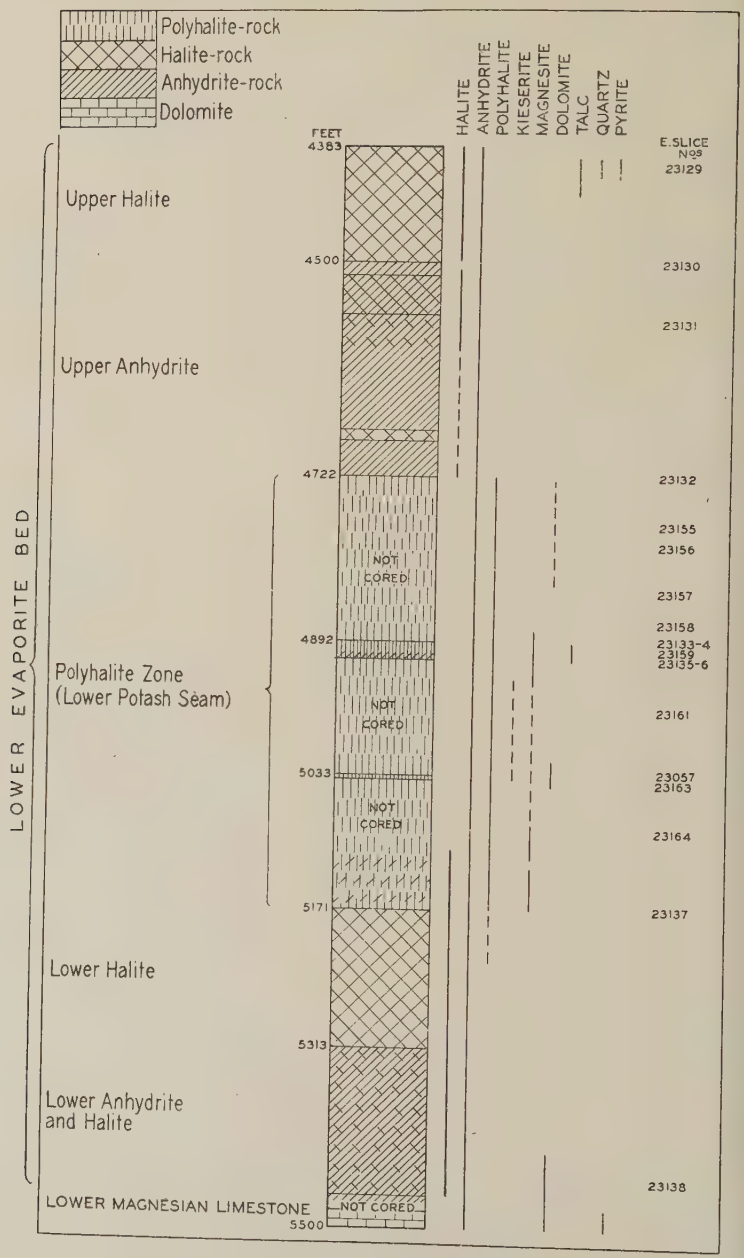


FIG. 1. Section of the lower evaporite bed of Eskdale no. 3 (Sleights) borehole, to show approximate ranges of minerals.

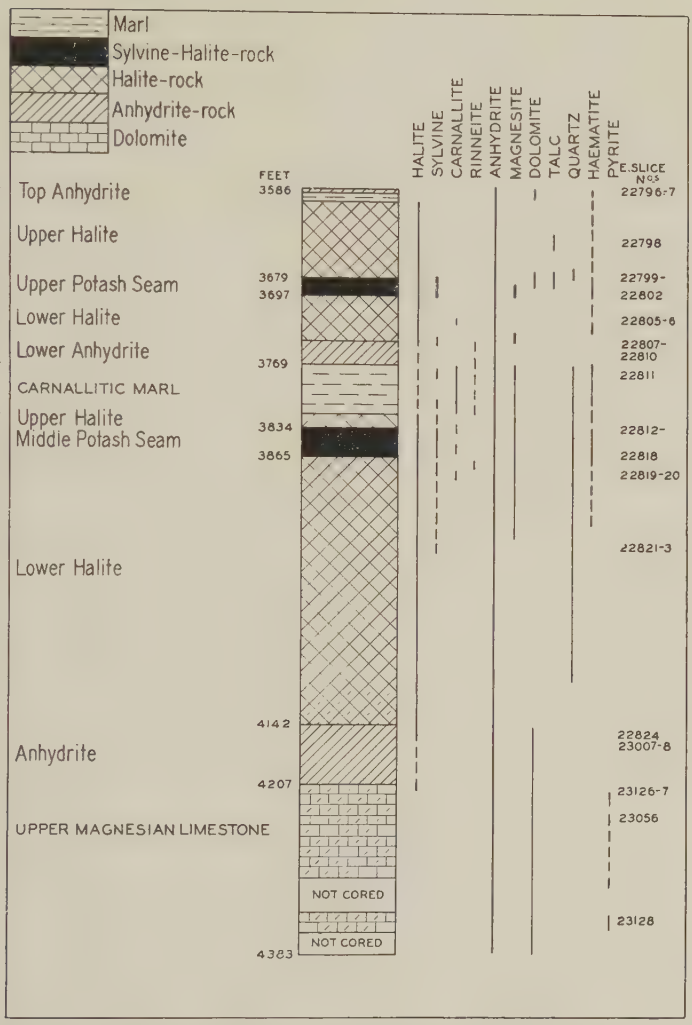


FIG. 2. Section of the upper Magnesian Limestone, middle and upper evaporite beds, and carnallitic marl at Eskdale no. 3 (Sleights) borehole, to show approximate ranges of minerals.

and X-ray investigations. Chemical investigations were the work of Mr. Harvey and Mr. Waters. The assistance of Dr. J. A. C. McClelland of the Government Laboratory, who made the spectrographical determinations, is gratefully acknowledged.

We are indebted to Mr. W. R. Madel and Dr. O. B. Westcott, successively in charge of the boring campaign under Mr. J. Hughes, Technical Manager of the Wilton Works of Imperial Chemical Industries, for much kindness; also to the late Mr. L. H. Barker, to Mr. B. T. James, as well as to Mr. L. R. Raymond and Mr. M. E. Pease, respectively geologist and mineralogist in the Research Department, I.C.I., Billingham.

The first published record of a potassium salt in Britain is due to C. E. Tilley (1943), who identified polyhalite in a core from Eskdale no. 2 boring submitted to him by W. G. Fearnsides. A log and description of this boring were published by G. M. Lees and A. H. Taitt (1945). Miss E. M. Guppy (1944), discovered boracite in cuttings from Eskdale no. 2 boring. More recently in a notable series of papers, F. H. Stewart (1949, 1951 *a*, *b*) has described the petrography and petrology of the evaporites. The results of our examination of the material from the more recent borings in a great measure confirm the conclusions reached by Stewart; attention is therefore concentrated here upon new features. These include (1) the identification of kieserite from the polyhalite zone; (2) a description of the rich middle sylvine zone, possibly represented by material lost in solution at Aislaby; (3) an account of the carnallite-bearing marl overlying this zone; (4) a description of the sylvine-halite seam in the upper evaporite bed.

The difficult task of making the 150 thin sections used in our investigations was carried out by Mr. D. Hepple; Mr. J. Rhodes took the photomicrographs.

THE PERMIAN EVAPORITE SUCCESSION.

The sub-divisions of the Permian evaporite succession in the Whitby area are tabulated below.

Upper evaporite bed 150–215 feet	Top anhydrite (p. 687) Upper halite (p. 687) Upper potash seam (p. 685) Lower halite (p. 684) Lower anhydrite (p. 684)
Carnallitic marl (p. 681) 40–55 feet	
Middle evaporite bed 239–383 feet	Upper halite (p. 681) Middle potash seam (p. 677) Lower halite (p. 676) Anhydrite and banded anhydrite-halite zones (p. 676)

Upper Magnesian Limestone
183 feet

Lower evaporite bed
1098 feet
- Upper halite (p. 675)
- Upper anhydrite (p. 674)
- Polyhalite and kieserite zone (lower potash seam) (p. 672)
- Lower halite (p. 672)
- Lower anhydrite and halite (p. 671)

Lower Magnesian Limestone

Lower Magnesian Limestone.

The lower Magnesian Limestone and the overlying lower evaporite bed was reached in only one of the three borings (Eskdale no. 3), and in these beds 'the rock was so hard that coring became too laborious and expensive to continue, and even straight drilling became a very slow process' (Fleck, 1950, p. s11) at the depth reached, 4383–5500 feet.

Some 22 feet of the lower Magnesian Limestone were penetrated at the bottom of the hole, and of this the last 4 feet were cored. The rock (E.23058)[1] is a buff laminated dolomite-rock containing discrete rhombs of dolomite and lenticles carrying dolomite, pyrite, and gypsum in a matrix of cryptocrystalline dolomite. It is uncertain whether the gypsum is of primary or secondary origin. Thin carbonaceous shaly partings occur in the rock.

Lower evaporite bed.

A comparison of the details of this bed in Eskdale no. 3, summarized by Fleck (1950, p. s10) with that given by Stewart (1949, p. 625) for Eskdale no. 2, reveals the following significant differences: (1) the total thickness is much greater in the recent boring (1098 feet—as compared with 460 feet; (2) the lower anhydrite in Eskdale no. 3 is poorly developed, its place being largely taken by halite-rich strata; (3) the polyhalite zone is more than five times as thick in Eskdale no. 3 as in Eskdale no. 2. The significance of the differences is not yet clear.

Lower anhydrite and halite.—Anhydrite-rock enclosing some halite occurs above the Magnesian Limestone; occasional black shaly films occur and, at 5465 feet depth, veinlets filled with halite traverse the rock. Occasional thin beds of fine-grained dolomite-rock are found; in one of these (E.23138) the carbonate, with ω 1·682, has a platy habit especially in coarser layers where it is associated with anhydrite. Between 5438 and 5464 feet depth, the anhydrite-dolomite-halite-rock

[1] Numbers in brackets refer to specimens in the sliced rock collection of the Geological Survey.

shows traces of distorted bedding. Above 5438 feet dark halite predominates up to 5313 feet, but an open mesh of anhydrite is present. It is probable, as Stewart has suggested, that this texture is inherited from the replacement of course primary gypsum by halite, the pseudomorphs being surrounded by anhydrite with a little dolomite.

Lower halite.—Above 5313 feet the cores up to 5171 feet consist essentially of halite having a predominantly grey colour, possibly due to carbonaceous inclusions. Halite crystals up to 2·5 cm. across are common. There are suggestions of gypsum pseudomorphs between 5250 and 5215 feet depth. A little polyhalite begins to appear towards the top of this zone, and at 5173 feet (E.23137) the presence of magnesite crystals about 0·005 mm. diameter was noted. These have centres, possibly of dolomite, of markedly lower refringence than the rest of the crystal. In the halite tiny hair-like needles, showing negative elongation, occur in bunches. It was considered possible that these might be the borate pinnoite, but their refractive index appears to be too high; further investigation is necessary.

Polyhalite and kieserite-bearing rocks (*lower potash seam*).—Of the 449 feet (5171–4722 feet depth) assigned to the polyhalite zone in Eskdale no. 3 boring, cores were taken for 21 feet only. Only one significant addition can be made to Stewart's account (1949) of the petrography of this zone: the presence of the mineral kieserite $MgSO_4.H_2O$, of which this constitutes the first record in Britain. The solubility of this mineral is such that it may well have been destroyed before the relevant portion of the core of the earlier boring was investigated.

The core from 5033 feet 6 inches depth in no. 3 boring (E.23057, 23057A) is a buff, semi-transparent rock with large white patches. The most abundant mineral present is polyhalite showing the characteristic polysynthetic twinning (see, for example, Shaller and Henderson, 1932, p. 72), optically negative with medium 2V, refractive indices α 1·550, β 1·562, γ 1·568. The polyhalite forms a mosaic of closely interlocked xenomorphic crystals. The white patches (pl. XXV, fig. 4), up to 2·5 cm. across, prove to be kieserite, optically positive with 2V near 60°, and high birefringence as indicated by the refractive indices α 1·524, γ near 1·580. Many cleavages are visible in thin section, and striking polysynthetic twinning is displayed. The large kieserite crystals are considered to be porphyroblasts, for at their margins they appear to have replaced polyhalite, and they enclose many relics of polyhalite, reminiscent of the anhydrite relics often found in gypsum. Trains of tiny hexagonal or rounded carbonate crystals occur in the rock, similar

those described from the main anhydrite and salt-field of SE. Durham (Dunham, 1948). Refractive index measurements indicate that both magnesite (with ω 1·703) and dolomite (ω 1·682) are present. Small pods containing halite, more coarsely crystallized than the polyhalite, and very little anhydrite are present, and these appear to have been affected by polyhalite. Thus in this rock the sequence of replacements seems to have been: anhydrite, halite → polyhalite → kieserite.

The presence of kieserite was confirmed by X-ray powder photography, the interplanar spacings computed agreeing well with those given for $MgSO_4.H_2O$ in the 'Index of the American Society for Testing Materials' (No. I–1580).

An analysis of the rock, with a calculated proximate mineral composition, is given in table I. Although a noteworthy amount of strontium is revealed, no separate strontium mineral was found, and it is surmised that this element may substitute for calcium in polyhalite.

TABLE I. Partial analysis of polyhalite-kieserite-rock. No. 3 (Sleights) boring, 5033 feet 6 inches.

Soluble in dilute HCl:		Determined spectrographically:	
Ca	10·95	Li	not detected, less than 0·01
Mg	5·36	Rb	,, ,, ,, 0·1
K	11·05	Cs	,, ,, ,, 0·1
SO_4	56·82	Ba	not detected
Cl	0·77	Pb	not detected
CO_3	3·63	Mn	slight trace
Total H_2O+105° C. ...	6·25	Sr	0·9
		B	0·02 (approximate)
	94·83	Tl	not detected

Proximate mineral composition:

Polyhalite (strontium-bearing)	$K_2Ca_2Mg(SO_4)_4.2H_2O$ ...	85·2
Kieserite	$MgSO_4.H_2O$	3·6
Magnesite	$MgCO_3$	4·4
Dolomite	$CaMg(CO_3)_2$	0·2
Halite (from excess chlorine)...	NaCl	1·3
Difference (siliceous matter) ...	—	5·3
		100·0

Analyst, W. F. Waters, Geological Survey and Museum, Lab. no. 1542 (1949). Spectrographical determinations by J. A. C. McClelland (1950).

The next higher core-samples were taken in the range 4892–4910 feet (E.23133–6). Here the rocks are greyish-white and contain in order of abundance, anhydrite, polyhalite, magnesite, dolomite, halite, and talc. The polyhalite forms a hypautomorphic-granular aggregate of elongate

crystals 0·01–0·06 mm. long, with, in places, larger individuals up to 2·0 mm. across. This aggregate appears to have invaded, mainly along vertical directions, somewhat coarser anhydrite-rock. Magnesite (ω 1·701) occurs as tiny hexagonal crystals, forming streaks in the rock. Some of these have centres of markedly lower refringence, which may be dolomite (E.23136). In some rocks, the anhydrite has probably recrystallized to form coarse poikilitic areas which may have formed contemporaneously with equally coarse polyhalite.

The continuity of the occurrence of polyhalite through the extensive uncored portion of the polyhalite zone was established by Imperial Chemical Industries' laboratories by the determination of potash-content in the drill-cuttings, both radiometric and chemical methods of analysis being employed. For our microscopical examination, cuttings were mounted in bakelite and sectioned, using the technique successfully developed at the Geological Museum in connexion with the D'Arcy Exploration Company's boring programme. The sections (E.23155–23164) show polyhalite, anhydrite, and magnesite in the fragments; the presence of clear primary kieserite was established in cuttings from 4946 feet, but some fine-grained, dusty-looking kieserite in specimens from other depths was considered to be either an alteration of polyhalite during drilling, or else a recrystallization product. All the fragments were more or less completely coated with secondary gypsum.

At the top of the polyhalite zone, at 4722 feet depth (E.23132) there occurs a rock rich in anhydrite which encloses lenticular wisps of fine-grained magnesite. Interlayered with the magnesite is a pale brown isotropic mineral, showing no cleavages and having refractive index appreciably below that of halite, and close to 1·530. It is believed that this may be langbeinite, $K_2SO_4.2MgSO_4$, but the very meagre quantity of the material precludes satisfactory identification. A careful watch for this mineral, which is abundant in the New Mexico potash-field, is desirable as investigation of the British field proceeds. Halite has to some extent replaced the anhydrite in this rock, and coarse polyhalite appears to have been the latest mineral to crystallize here. X-ray data on polyhalite are given on p. 688.

Upper anhydrite.—Predominantly anhydrite-rocks, with some halite, especially near the top, range from 4722 up to 4556 feet. At 4566 feet (E.23131) lenticular masses of anhydrite are bordered with very fine-grained carbonate; here there is a suggestion of dips up to 20°. At 4637 feet finely banded white anhydrite seems to have been folded, contorted, and invaded by halite. A horizontal lineation appears in a

bed of grey halite between 4676 and 4684½ feet, beneath which apparently contorted anhydrite is again seen. The possibility of movement involving flowage of halite may perhaps be envisaged at this level.

Upper halite.—The topmost division of the lower evaporite consists of banded halite-anhydrite-rock overlain by massive halite. A crude lineated or schistose texture is seen in places between 4423½ and 4500 feet, while between 4530 and 4556 feet the halite encloses apparently broken-up anhydrite-rock fragments showing dips up to 25°. At higher levels clear rock-salt shows thin bands of dark pigmentation dipping at about 20°. Talc is a prominent constituent at 4410¼ feet and again at 4403 feet, as silvery-looking plates up to 3·5 mm. across. The other constituents at 4403 feet (E.23129) are halite, anhydrite, scarce euhedral quartz crystals, and pyrite. The talc lies, as Stewart has described, along cleavages in the halite, and some crystals also occur along what may be bedding-planes in the rock.

Upper Magnesian Limestone.

Dolomite reappears in force at 4383 feet, and dolomite-rock extends up to 4207 feet. In places the rock is shaly, with layers containing clay minerals and carbonaceous matter. Anhydrite is generally present in small amount and usually with poikiloblastic habit or in a form indicative of porphyroblastic growth. Pyrite is also widespread in small amounts, in places edging the anhydrite areas. Remains of the fossil *Filograna ?permiana* (King) were found at 4237–4248 feet, and again at the top of the division, from 4207–4210 feet (E.23126). The remains consist of tubes up to 0·75 mm. across, and up to 3·0 mm. long. In cross-section they now consist of alternating layers of dolomite and anhydrite, giving evidence of former metasomatic changes (Dunham, 1948, pl. X, fig. 3). This fossil[1] has been accepted elsewhere as evidence of an Upper Zechstein age (C. J. Stubblefield, in discussion on Dunham and Rose, 1949, p. 40). The identification in this case was made by Dr. Stubblefield (E.18374).

Middle evaporite bed.

This bed was penetrated in all three borings; it was completely cored, with excellent recovery, in Eskdale no. 3 boring, while in Eskdale no. 4 the coring was discontinued before the basal anhydritic zone was reached. In no. 6 borehole the base of the bed was not reached.

[1] The fossil also occurs in Eskdale no. 2 borehole at a similar horizon, at 4218 feet depth.

Anhydritic and banded anhydrite-halite zones.—In no. 3 boring anhydrite, with dolomite and occasional low-dipping, black, shaly partings, extends from 4207 up to 4195 feet. Above this, layers not more than a few inches thick of anhydrite and halite alternate up to 4184 feet. It is tempting to regard these as 'annual' layers such as have been recorded from the German evaporite field, in which anhydrite is believed to have been deposited during the warmer season and halite during the cooler part of the year, in conformity with temperature-coefficients of solution of the minerals (J. D'Ans, 1949). Whether this is so or not, the rocks almost certainly represent primary precipitates. Evidence of disturbance of anhydrite layers and their enclosure in halite (probably recrystallized) is found at intervals from 4184 up to 4164 feet; dips in anhydrite-rock fragments as high as 20° may be seen in this range. Between 4164 and 4162½ feet a buff-grey porcellanous rock (E.23007–8) is a dolomite-siltstone in which the carbonate averages 0·005 mm. in grain-size, and the rock encloses a few detrital quartz grains. In section, groups of rounded areas, filled with anhydrite or halite, suggest the possible former presence of a fossil such as *Filograna*. Wavy markings traversing the rock may be the remains of ripple-marks.

At slightly higher levels, arborescent growths of anhydrite in halite occur, a texture probably inherited from the replacement of early gypsum (see Stewart, 1951*a*, for a full discussion of comparable textures at similar levels in Eskdale no. 2 boring). Banded rocks continue up to 4114 feet, but the halite bands in the upper part of this zone are markedly thicker than those of anhydrite.

Lower halite.—This zone ranges from 4114 up to 3865 feet in no. 3 boring, where it was cored throughout. In no. 4 boring the range was 4386 up to 4245 feet 10 inches, cores being taken only above 4330 feet. The zone was penetrated in no. 6 borehole from 4055 down to 4215 feet, below which anhydrite bands were found.

Halite is the dominant mineral, with anhydrite as a persistent accessory. In the lower part of this zone as exposed by no. 3 boring, the halite is nearly colourless, or has a greyish shade, probably due to small amounts of carbonaceous pigment. Higher up, the salt is pink, with, rarely, a few bright red individual halite crystals (e.g. at 4036 feet in Eskdale no. 3 boring). A few partings of dark shaly material occur apparently undisturbed.

Automorphic quartz crystals make their appearance, as in the Aislaby boring (Stewart, 1951*a*, p. 451) low down in this zone, at 4000 feet depth in no. 3 boring, and occur in all the powders and slices examined up to

s summit. Cuttings from no. 4 boring at the base of the zone, some of hich probably came from higher levels (E.23704) include quartz thedra up to 1·0×0·2 mm., fragments of quartzite-like rock composed clear interlocked hypautomorphic quartz crystals enclosing anhydrite ystals, and also fragments showing the poikilitic enclosure of magnesite d anhydrite by quartz.

Magnesite, as automorphic hexagonal plates, flattened parallel to 001), and with ω 1·701 1·703, first appears, according to our observaons, at 3952 feet in no. 3 boring and at 4300 feet in no. 4. It continues oradically to the top of the zone.

From 4000 feet depth in no. 3 boring upwards, occasional occurrences sylvine have been noted. Owing to the greater solubility of the otassium chloride, the former presence of sylvine is revealed by conuous deep etching on the outsides of the cores. The mineral occupies egular gashes and pockets, and this mode of occurrence, in conjuncon with microscopical evidence of replacement of halite by sylvine .22819–23) makes it likely that potassium penetrated into this zone eneath the rich middle seam. Carnallite occurs as rounded crystals or roups of crystals enclosed in halite (E.23701–3). These have the appearnce of primary constituents, but, as the extent to which the salt has reystallized is uncertain, such appearances must be treated with caution.

Middle potash seam.—The paragenesis sylvine-halite-magnesite-uartz-anhydrite-(carnallite), which first makes its appearance in the alite zone, reaches its maximum development in the middle potash am. The depth-ranges of this seam have been given by Fleck (1950, g. 10): 3865–3834 feet in no. 3 boring; 4245 feet 10 inches–4213 feet no. 4 boring; 4055–4043 feet in no. 6 boring. Sylvine occurs: (1) as d crystals wholly pigmented with haematite inclusions in the form of ods and hexagons, individually visible in many specimens only at the ghest magnifications; (2) as crystals marginally red, but with white ntres; (3) as milky-white crystals. In thin section the sylvine is adily distinguished from halite by its much lower refractive index ·490, from which no detectable variation was found in numerous terminations). Smoothly curving boundaries separate the sylvine from djacent halite. Much of the middle seam as exposed in no. 3 boring onsists of a phanerocrystalline aggregate of red sylvine intergrown with alite, which is mainly dark grey probably due to associated carbonaeous material. On the other hand, red pigmentation is conspicuously cking from most of the sylvine in this seam in no. 4 boring. The grainze in both cases is coarse, individual crystals reaching 1–3 cm. across.

Enclosed in the chlorides there are automorphic anhydrite crystals of very variable size, and euhedra of magnesite and quartz. Some clay is generally present, and in some, though not all, cases quartz and magnesite crystals cluster round the clay inclusions, or occur with clay which occupies an interstitial position among the chlorides. A fringe of haematite pigmentation in chlorides sometimes surrounds clay inclusions, as if ferric oxide had diffused from the marl into the chlorides. Such pigmentation is not capable of resolution into separate haematite crystals, even at the highest magnifications.

Carnallite is probably not much more abundant in the seam than in the upper part of the halite zone, and its mode of occurrence is similar, generally as rounded or crudely hexagonal inclusions in halite. On cores from the top of the seam in no. 3 boring, an efflorescence which formed on the outside was found to consist largely of carnallite. This may be the product of reactions involving the boring fluid, or carnallite may be soluble in the fluid, but capable of redeposition.

From no. 3 boring a sample was taken for chemical analysis in Geological Survey and Museum laboratories, the first object being to investigate the bromine- and iodine-contents; subsequently full analyses were made of water-soluble and insoluble portions. The sample consisted of small chips removed from the core of the seam at regular intervals of 3 inches, the partially dissolved outside of the core being avoided in taking the chips, or removed from them. This method of sampling is less satisfactory than that later adopted by I.C.I. for sampling the cores of no. 4 and subsequent borings, when a diamond-impregnated cutting disk was used to make a longitudinal cut. This explains some discrepancies between the result obtained by I.C.I. and by us, to which Fleck (1950, p. s11) makes reference. The results of the Geological Survey analysis are given in table II, together with one possible calculation—not, of course, a unique solution—of the mineral composition.

Bromine was found to be present in the water-soluble portion of the core, but iodine was not detected, the method used being sensitive to 0·001 %. Comparison of the bromine-content of the middle seam (0·104 %) with that of the upper seam (0·014 %, p. 686) suggests a correlation with total KCl-content rather than with amount of carnallite present, which is approximately the same in both seams. This may, however, be accidental, and in calculating the mineral composition it has been assumed that the bromine occurs as a substitution for chlorine in carnallite.

TABLE II. Analysis of middle potash seam.

No. 3 (Sleights) boring, 3834–3865 feet.

Water-soluble matter.		Water-insoluble matter.	I.	II.
Mg	0·09	SiO_2	51·8	2·36
Ca	1·22	Al_2O_3	13·0	0·59
Na	19·47	Fe_2O_3	2·7	0·12
K	21·70	FeO	2·0	0·09
SO_4	2·74	MgO	15·7	0·72
Cl	49·83	CaO	0·4	0·02
Br	0·104	Na_2O	trace	trace
Li*	n.d. < 0·01	K_2O	0·8	0·04
Rb, Cs*	n.d. < 0·1	ign.†	8·9	0·41
Pb*	trace	TiO_2	0·8	0·04
Ba*	n.d.	CO_2	2·4	0·11
Sr*	trace	SO_3	0·6	0·03
	95·15	FeS_2	0·9	0·04
Water-insoluble matter	4·57	Cr_2O_3	n.d.	n.d.
Total $H_2O-105°$ C. ...	0·25	Li_2O	n.d.	n.d.
	99·97	P_2O_5 (qual.) ...	trace	trace
Proximate mineral composition:		Cl (qual.) ...	trace	trace
Halite	49·5	B_2O_3*	about 0·1	about 0·005
Sylvine	40·7	V_2O_3*	trace	trace
Carnallite	1·0	BaO*	trace	trace
Anhydrite	4·0	SrO*	trace	trace
Magnesite	0·2		100·1	4·57
Talc	2·0			
Ilmenite	0·08			
Haematite	0·1			
Pyrite	0·04			
Silicate minerals ...	1·9			
	99·5			

I. Water-insoluble matter as analysed.

II. Water-insoluble matter recalculated to 4·57%.

† Matter volatile on ignition, other than CO_2 and S, mainly $H_2O + 105°$ C.

The water-insoluble matter was analysed by semi-micro methods. Analyst, C. O. Harvey, Geol. Surv. Lab. nos. 1539, 1540 (1949).

* Semi-quantitative spectrographic determinations by J. A. C. McClelland. n.d., not detected.

The cores showed no visible fluorescence when exposed to ultra-violet rays of 'long' or 'short' wave-length. In this connexion the absence of manganese is noteworthy, for Murata and Smith (1946, p. 527) state that halite containing small amounts of both Mn and Pb shows red fluorescence. Thallium, also an activator of KCl (Zuk, 1946), was not detected.

Trace elements likely to be present in hydrolitic products are the elements of intermediate 'ionic potential'. Dr. McClelland examined

the insolubles both from this seam and the upper seam spectrographically, and found no indication of the following elements: Sb, Zr, Nb, Ta, W, Ga, Ge, In, Yt, Eu, La, Ce, Th, Ni, Mo, Bi, Be.

Both in the halite zone and the middle seam a search was made for boracite, but without success. It will be recalled that Miss Guppy's (1944) discovery of this mineral was in cuttings, the residue from extensive solution of chlorides. Examination by Dr. McClelland of insoluble residues from five samples from no. 3 boring at the depths shown below gave the following approximate boron-contents:

Depth, feet.	B %.
4036	0·01–0·05
4046	trace
4090	0·05–0·10
4110	0·01–0·05
4131	0·01–0·05

In the insolubles from the middle seam the boron was of the order of 0·1 %. Though a concentration of boracite in the middle seam, such as Stewart thought might be possible (1951 *a*, p. 449), has not been confirmed, a little more boron appears to be present than in the halite zone.

The middle seam as exposed by no. 4 (Sneaton) boring shows alternations of halite-rich and sylvine-rich layers. It is strikingly different in appearance from the Sleights material, being predominantly white or grey rather than red and black. Fleck (1950, p. s11) has given the KCl-content of the various beds. In the table below, proximate mineral compositions, based by kind permission on I.C.I. analyses, are given:

Proximate mineral compositions, middle seam, no. 4 (Sneaton) boring.

Depth. (Feet/inches).	Thickness. (Inches).	Halite.	Sylvine.	Carnallite.	Insoluble.
4213–4214/2	14	59	37	1	1%
4214/2–4215/2	12	78	2	5	10
4215/2–4217/3	25	38	58	1	1
4217/3–4221	45	50	4	5	33
4221–4230/11½	119½	58	38	1	2
4230/11½–4241	120½	95	2	1	1
4241–4245/10	58	57	38	1	2

The next 96 inches below the seam showed 96 % halite; sylvine and carnallite 1 % each. The high carnallite-contents of the fourth and sixth beds of the seam are confirmed by microscopical examination. The mineral occurs in close association with clay present as thin beds or inclusions in the seam.

In no. 6 (Upgang) borehole the middle seam is only 12 feet in thickness and contains greater quantities of marl than in boreholes E3 and E4. Rinneite occurs at 4047 feet and carnallite at 4052 feet depth.

Alternations of sylvine-rich layers are present as in borehole no. 4. Proximate mineral compositions, based by kind permission on I.C.I. analyses are given below. Complete recovery of the core was not obtained and precise depths are not therefore stated.

Proximate mineral compositions, middle seam, no. 6 (Upgang) boring.

Portion of core 4041–4051 feet.	Halite.	Sylvine.	Carnallite.	Insoluble.
First 17 inches ...	75	< 1	3	19%
Next 7 „ ...	84	2	1	4
„ 11 „ ...	67	6	5	15
„ 35 „ ...	41	42	1	10
„ 10½ „ ...	32	19	1	34
„ 13 „ ...	33	37	3	17
„ 9 „ ...	40	17	1	28
4051–4062 feet.				
First 17 inches ...	54	31	1	7

The Geological Survey Museum slices of material from the middle seam include the following: no. 3 borehole E.22812-8; no. 4 borehole E.23694–23701; no. 6 borehole E.24292–7.

Upper halite.—In nos. 3 and 6 borings, the middle potash seam is overlain by a thin bed of halite, almost colourless, but enclosing some black and green marl and traces of bright red sylvine. There is a suggestion of veining or steep banding at this level in the Sleights core (3821–3834 feet). In no. 4 boring, the middle seam is overlain directly by the carnallitic marl.

Carnallitic marl.

In no. 2 (Aislaby) boring, a bed of marl was recorded above the middle evaporite bed, at 3861 3920 feet. This bed, which has proved to be a persistent one, shows some unusual features in the more recent borings. Fleck (1950, pp. s9, s11) has recorded that the presence of carnallite was established here both by I.C.I. laboratories and at the Geological Museum. More recently a specimen of the marl from Messrs. Fisons' Robin Hood's Bay no. 1 boring has been exhibited by Hollingworth and Stewart (1950).

In no. 3 boring recovery was poor, only centres of cores, partly fragmented, being obtained with a mass of loose material. Carnallite, as grains with low refringence but high birefringence (α 1·468, γ 1·493,

optically positive, 2V near 70°, occasionally showing polysynthetic twinning) proved to be common among the marl fragments, accompanied by halite, anhydrite, quartz, haematite, and clay minerals (E. 22811). The presence of carnallite as a main constituent was considered to have been the reason for the disintegration of the marl during drilling. The occurrence of rinneite was also suspected. This mineral, first found by Stewart (1949) in the upper evaporite bed at Aislaby, yields on decomposition sylvine and limonite of a conspicuous yellow colour. Such material was first noted in the marl at 3769 feet by Mr. L. R. Raymond.

In no. 4 boring the recovery from this bed was again poor, but here an average sample covering the range 4173–4213 feet was taken for analysis in the Geological Survey laboratory. The results are given in table III. There are too many uncertainties to make possible a calculation of the proximate mineral composition at this stage. It is, however, possible to state that the carnallite content is of the order of 10 % (a figure corroborated by later I.C.I. analyses on cores from no. 6 boring) and that sylvine amounts to about 0·4 % and halite 17·0 %. Among the insoluble constituents, magnesite probably amounts to 2·7 %, but a large residue of magnesia (5·28 %) is left unaccounted for. As in the case of the insolubles in the middle potash seam, magnesia is exceptionally high, implying, it is suggested, the presence of talc or magnesian chlorite or both. Among the clay constituents, the potash figure is normal and no doubt represents white mica, probably with an illite-type mineral. The soda could be accounted for by montmorillonite. Considerable amounts of kaolin must be present.

In no. 6 boring much greater success was achieved with coring the carnallitic marl. The rock is a consolidated salt-clay, dull red except near the base, where the coloration changes to yellowish or olive-green. In this section (E.24103–5, 24112–27) it is seen to be full of tiny ramifying veinlets, some more or less horizontal, spaced at intervals of 0·05–1·0 mm. apart, linked by steep, nearly vertical veinlets. These veinlets seldom exceed 0·1 mm. in width.

Oblique veins, up to 3 cm. wide, carrying pink and white salts also occur, though at wide intervals. The minerals in the largest veins include halite, carnallite, sylvine, and rinneite, with in places some anhydrite, euhedral quartz, and magnesite (pl. XXV, fig. 3). In the tiny veinlets, carnallite seems to be the dominant mineral, but determination is difficult in the smallest ones. The clay also encloses cubic pseudomorphs after halite—now partly replaced by or occupied with quartz

TABLE III. Analysis of carnallitic marl, No. 4 (Sneaton) boring. 4173–4213 feet.

Water-soluble matter.		Water-insoluble matter.	I.	II.
g	0·88	SiO_2	56·05	35·83
a	1·99	Al_2O_3	16·37	10·47
a	6·72	Fe_2O_3	3·54	2·28
... ...	1·62	FeO	1·16	0·74
O_4	2·77	MgO	10·32	6·59
l	15·69	CaO	0·33	0·21
r	0·041	Na_2O	0·26	0·17
... ... n.d. <	0·001	K_2O	2·70	1·73
n*	trace	ign.†	5·67	3·62
O_2	n.d.	TiO_2	0·82	0·52
*	n.d.	P_2O_5	0·15	0·10
a*	n.d.	CO_2	2·20	1·42
*	trace	SO_3	0·22	0·14
*	trace	B_2O_3*	about 0·2	about 0·1
*	trace			
b*, Cs*	n.d.	FeS_2	0·06	0·04
*	n.d.	Cr_2O_3	0·02	0·01
		BaO	0·03	0·02
	29·71	SrO	trace	trace
Water-insoluble matter	64·0	Li_2O	0·03	0·02
Total $H_2O-105°$ C. ...	5·6		100·13	64·0
	99·3			

I. Water-insoluble matter as analysed.
II. Water-insoluble matter recalculated to 64·0 %.

The two portions were analysed separately; analysts, W. F. Waters and K. L. H. urray, Geol. Surv. Lab. nos. 1563, 1564 (1950).
* Semi-quantitative spectrographic determination by J. A. C. McClelland.
n.d., not detected.
† Matter volatile on ignition, other than CO_2 and S; mainly $H_2O+105°$ C.

E.24126)—and pseudomorphs showing the characteristic outline of ypsum, replaced by halite and anhydrite (E.24115–17). From the fact nat carnallite veinlets follow the sides of the halite pseudomorphs, it concluded that introduction of carnallite occurred after their replace- ent (E.24127). Polyhalite occurs in the marl in the form of small ounded aggregates, partly coated with clay (E.24113, 24125).

The clay minerals are too much mixed with haematite to permit ptical study except in the basal layers of the bed. Here attempts to etermine the refractive indices suggested that these are abnormally w, but the presence of secondary fine-grained gypsum complicates the

investigation of this material. The possibility that montmorillonite might be present (Fleck 1950, p. s12) was suggested to explain the swelling and breaking-up of the clay in water, but X-ray determinations of glycerol-treated material have failed to substantiate the presence of a mineral of this group. Possibly the breaking-up can be explained sufficiently by the solubility and deliquescence of carnallite and rinneite. Examination by X-ray powder methods, after repeated extraction of soluble salts with water, of a specimen from the carnallite marl of borehole E6 (4016 feet depth) shows the clay fraction to consist of quartz and illite with possibly a member of the kaolinite group.

This clay formation, as seen at the present early stage of investigation, has evidently had a complex history. Gypsum and halite, and perhaps also polyhalite, seem to have been deposited with the clastic constituents: later brines brought in anhydrite, quartz, and further halite. Subsequently carnallite and rinneite were introduced along veins which show evidence of still later attack by anhydrite, halite, sylvine, quartz, and magnesite.

Upper evaporite bed.

Lower anhydrite.—The uppermost evaporite sequence is initiated by anhydrite with some layers of dolomite containing angular detrital quartz grains (e.g. 3975 feet in no. 6 boring, E.24111). Pink halite is sporadically present, occurring in the form of striking pseudomorphs after early gypsum (E.24109–10), such as Stewart has described. These occur in a matrix of anhydrite which appears, from the vagueness of the halite-anhydrite margins, to have crystallized during or after the metasomatism of the gypsum. Veins carrying halite, sylvine, and yellow limonite (probably after rinneite) cut through the anhydrite, for example at 3752 feet in no. 3 boring. The anhydrite zone is very consistent in the three borings, the thicknesses being respectively 26, 28, and 26 feet. Towards the top of the lower anhydrite in both no. 3 and no. 6 boreholes, there is a subsidiary development of sylvine. This sylvine (pl. XXV, fig. 5) is often pigmented by haematite, contains automorphic crystals of anhydrite, and displays sinuous boundaries against associated halite, e.g. from 3957 to 3959 feet depth in Eskdale no. 6 (E.24298; E.24109).

Lower halite.—Above the anhydrite, flesh-pink halite, corresponding to Stewart's 'granular halite' zone, appears. A little anhydrite is mixed with it in the form of oblique wisps and veinlets. In the main this salt is unusually pure, but at its base there is a continuation of the sylvine

one which develops towards the top of the underlying anhydrite. This lvine zone ranges up to 3724 feet in Eskdale no. 3, and cores from this orehole show a strong yellow coloration of the sylvine at 3766 3769 feet epth. Although no rinneite was found here, it seems certain that this ariety of sylvine arises from a decomposition of rinneite into sylvine, alite, and yellow limonite. In no. 6 borehole at 3953 feet 3 inches to 955 feet 5 inches this sylvine zone is represented by a bed of coarse ear granular halite with anhydrite inclusions and opalescent sylvine ystals, and includes a 4-inch band of pure sylvine. In Eskdale no. 4 ere are traces of sylvine at 4144–5 feet immediately above the lower hydrite.

Upper potash seam. –The position of this seam appears to be between ewart's 'granular halite' and halite zones, but it does not seem to be resented in the Aislaby boring. It represents a recrudescence of the neral assemblage which first appears in the middle seam. Sylvine, alite, anhydrite, quartz, and subordinate carnallite are the principal nstituents. Three important features distinguish the upper from the iddle seam: (1) the presence of substantial amounts of salt-clay inter-ixed with the upper seam; (2) the abundance of coarse talc as a replace-ent of halite in the upper half of the seam; (3) the paucity of magnesite it. Texturally the two seams have, nevertheless, much in common.

A general sample, taken by the same method as the analysed sample the middle seam (p. 679), gave on analysis the results in table IV.

Thin sections in the Geological Survey collection of the upper potash ams include the following: no. 3 boring, E.22801–4; no. 4, E.23688 -91; o. 6, E.24128–38. Red pigmentation of the sylvine, marginal or com-lete, is displayed, and it is noteworthy that the colouring matter in djacent halite has a brown or greyish-brown appearance by comparison. he occurrence of red pigmentation along the courses of veinlets filled ith talc may indicate that some of the colouring is secondary (E.24132). he clay occurs as apparently broken up fragments, and is also found the interstices between chloride crystals, which have evidently been ble to push it aside.

It should be noted here that no potash minerals are present at this osition in the Aislaby borehole (Eskdale no. 2) where, from general ratigraphic and thickness considerations, the upper potash seam might e expected to occur. At this position, talc, the presence of which is a ature of the upper potash seam in the later borings, occurs in associa-on with halite at 3778–3782 feet depth. The first appearance of potash the Aislaby borehole is at 3798 feet depth where colourless halite

TABLE IV. Analysis of the upper potash seam.
No. 3 (Sleights) boring, 3679–3697 feet.

Water-soluble matter.

Mg	0·09
Ca	0·82
Na	29·16
K	6·73
SO_4	1·81
Cl	50·94
Br	0·014
I ... ... n.d. <	0·001
P_2O_5	n.d.
Mn*	n.d.
CO_2	n.d.
B*	n.d.
Ba*	n.d.
Sr*	trace
Pb*	trace
Li* ... ... n.d. <	0·01
Rb*, Cs* ... n.d. <	0·1
Tl*	n.d.
	89·56
Water-insoluble matter	10·24
Total $H_2O+105°$ C. ...	0·25
	100·05

Water-insoluble matter.

	I.	II.
SiO_2	57·8	5·91
Al_2O_3	14·6	1·49
Fe_2O_3	3·3	0·34
FeO	1·9	0·19
MgO	9·7	0·99
CaO	0·2	0·02
Na_2O	0·2	0·02
K_2O	5·9	0·60
ign.†	5·0	0·51
TiO_2	0·8	0·08
P_2O_5 (qual.) ...	trace	trace
MnO*	trace	trace
CO_2	0·9	0·09
B_2O_3*	trace	trace
SO_3	trace	trace
Cl (qual.) ...	trace	trace
FeS_2	trace	trace
Cr_2O_3	n.d.	n.d.
V_2O_3*	trace	trace
BaO*	trace	trace
SrO*	trace	trace
Li_2O	n.d.	n.d.
	100·3	10·24

Proximate mineral composition:

Halite	74·1
Sylvine	12·5
Carnallite	1·0
Anhydrite	2·6
Magnesite, dolomite ...	0·1
Talc	2·9
Silicate minerals ...	6·8
Ilmenite	0·1
Haematite	0·3
	100·4

I. Water-insoluble matter as analysed.
II. Water-insoluble matter recalculated to 10·24%.
† Matter volatile on ignition, other than CO_2 and S, mainly $H_2O+105°$ C.

The water-insoluble matter was analysed separately by semi-micro methods. Analyst, C. O. Harvey, Geol. Surv. Lab. nos. 1537, 1538 (1949).
* Semi-quantitative spectrographic determination by J. A. C. McClelland.
n.d., not detected.

contains a small proportion of pink sylvine. At 3804–3805 feet the core is remarkable for the presence of pockets of a bright yellow mineral which proves to be sylvine mixed with finely divided limonite, which are probably the decomposition products of rinneite. Thus the potash zone in the upper evaporite bed at the Aislaby borehole bears a marked

imilarity and has a comparable position to the subsidiary potash zone t the base of the lower halite in the upper evaporite bed at Eskdale no. 3 borehole. It is considered possible, therefore, that the upper seam has died out completely before reaching the site of the Aislaby borehole, or else that it is present in the uncored portion of the hole between 3653 and 3738 feet.

Upper halite.—Halite, without potassium minerals, but enclosing streaks and layers of clay, is the next upward division in all three borings. Anhydrite is present in variable amount with, towards the top, some dolomite. The clay is the dull red type found also in the saliferous marl which forms the uppermost division of the Permian here, and occasional bluish or greenish 'eyes' are to be seen in it, where it is included in the halite. Some 'schistose' halite was noted at 3607 feet in no. 3 boring. Talc continues in small amount in this zone (E.22798).

Top anhydrite.—The topmost bed of the evaporite succession is a thin but persistent layer of anhydrite, composed of interlocked automorphic and hypautomorphic crystals averaging about 0·05 mm. Thin red films of haematitic clay enclose lenses of still finer-grained anhydrite which are engulfed within the main mass of the rock (E.22796).

The saliferous marl, mainly an anhydritic siltstone with detrital micas, quartz, clay minerals, and fine-grained dolomite, immediately succeeds the anhydrite (E.22794–5, 23686–7).

X-ray data.

As X-ray powder data for carnallite do not appear to have been published, a standard photograph, details of which are given (table VI, no. 1) of Stassfurt material, was prepared. In a number of photographs sylvine was always found to be present as an impurity and is therefore also shown in the table (no. 2). It should be noted that the relative intensities of sylvine and carnallite are likely to vary. In the Stassfurt specimen the ratio of the intensities of the strongest lines of carnallite and sylvine was 1:1·3. Powder data for polyhalite are shown in table VI, no. 3. Intensities expressed numerically in the table were measured with a microphotometer and are given relative to unit intensity for the strongest line.

Summary.

1. An account is given of the Permian evaporite succession revealed by the three Imperial Chemical Industries borings, with ranges of minerals as determined by optical methods. A lower polyhalite zone and one subsidiary and two main sylvine beds are described.

TABLE VI. X-ray powder data for evaporite minerals.
(λ = 1·5418 Å., 9-cm. camera.)

1, Carnallite.		2, Sylvine.		3, Polyhalite.			
d(Å.)	*I.*	*d*(Å.)	*I.*	*d*(Å.)	*I.*	*d*(Å.)	*I.*
4·65	0·5	3·13	vs	5·94	0·18	1·693	0·11
3·77	0·5	2·21	ms	4·77	< 0·05	1·584	0·06
3·56	0·5	1·808	mw	4·08	0·06	1·550	0·05
3·30	1·0	1·568	w	3·78	< 0·05	1·510	0·05
3·01	0·30	1·392	mw	3·47	< 0·05	1·453	0·10
2·92	0·7	1·282	w	3·39	0·19	1·434	< 0·05
2·83	0·33	1·108	vw	3·17	0·9	1·403	< 0·05
2·37	0·23	1·047	vw	2·90	1·0	1·377	< 0·05
2·33	0·31			2·54	< 0·05	1·354	0·06
2·12	0·27			2·50	< 0·05	1·327	< 0·05
2·00	0·34			2·41	< 0·05	1·303	0·05
1·975	0·43			2·33	0·05	1·267	< 0·05
1·886	0·10			2·21	0·17	1·231	0·08
1·865	0·11			2·03	< 0·05	1·215	< 0·05
1·458	0·20			1·984	0·09	1·197	< 0·05
				1·936	0·09	1·179	< 0·05
				1·891	0·12	1·152	< 0·05
				1·861	0·12	1·112	0·07
				1·805	0·15	1·059	0·06
				1·766	0·09	1·041	< 0·05
				1·725	0·08	1·021	< 0·05

1, Carnallite, Stassfurt, Germany (Film no. X55).
2, Sylvine, occurring with carnallite, Stassfurt (X55).
3, Polyhalite, Eskdale no. 2 borehole, Yorkshire (X147).

2. Kieserite accompanies polyhalite in the lower potash seam, where it appears to have been the last mineral formed, replacing polyhalite which had itself replaced anhydrite.

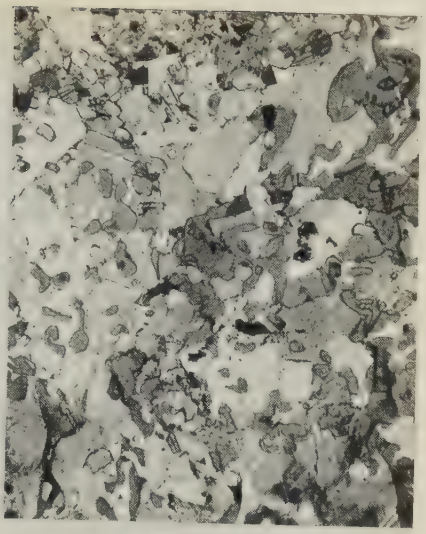
3. In the middle evaporite bed, the mineral assemblage in the middle potash seam includes sylvine, halite, minor amounts of carnallite, anhydrite, euhedral quartz, and magnesite. Chemical analysis reveals an abundance of magnesia in the water-insoluble portion of this seam.

4. The carnallitic marl is a salt-clay containing pseudomorphs after gypsum and halite. It is traversed by veins and veinlets carrying carnallite, rinneite, halite, sylvine, anhydrite, quartz, and magnesite. A little polyhalite, possibly of early formation, occurs in the clay. Carnallite forms 10 % of the rock.

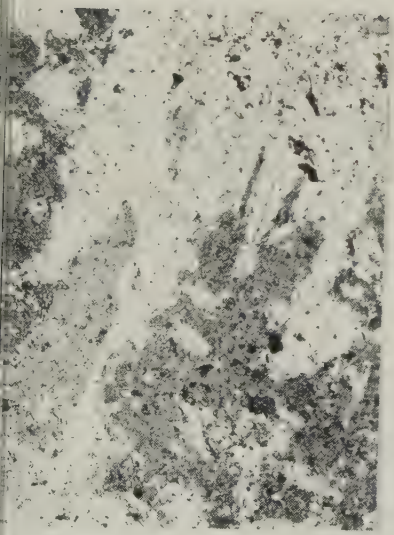
5. The upper potash seam, at or near the junction of Stewart's 'granular halite' (with scarce rinneite) and halite zones, contains halite, sylvine, subordinate carnallite, anhydrite, and quartz. Talc is abundant in the upper part of the seam.

6. The water-insoluble portions in analysed samples from the middle

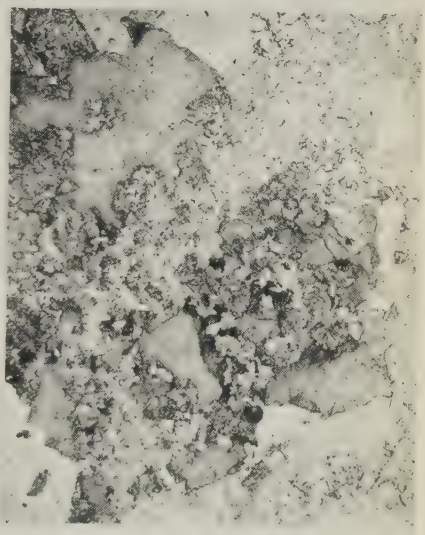
3a



3b



4



5

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and upper seams and from the carnallitic marl are alike in the abnormally high MgO/CO_2 ratio they show. This points to the presence of magnesium silicates such as talc (which is conspicuous in the upper seam) or magnesian chlorite or both.

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EXPLANATION OF PLATE XXV.

Photomicrographs of potash-salt rocks from NE. Yorkshire.

FIG. 3. Carnallite vein. Carnallite is distinguished by intersecting sets of twin lamellae. Halite shows sinuous boundaries against the carnallite. Small anhydrite euhedra occur at the top of the photograph. Eskdale no. 6 borehole, 4000 feet, E.24114. (*a*) Ordinary light, (*b*) nicols slightly uncrossed. ×12.

FIG. 4. Kieserite-polyhalite-rock. Kieserite, showing complicated twinning, encloses abundant grains of polyhalite. Eskdale no. 3 borehole, 5033 feet 6 inches, E.23057. Crossed nicols. ×57.

FIG. 5. Sylvine-anhydrite-halite-rock. Sylvine (cloudy) and subordinate halite (clear) enclose abundant small automorphic prisms of anhydrite. Eskdale no. 6 borehole, 3957 feet, E.24298. Ordinary light. ×14.

The clinopyroxenes of the Skaergaard intrusion, eastern Greenland.

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I. Introduction.

During the past twenty years considerable attention has been given to the compositions of the pyroxenes from basic igneous rocks. Hess (1941) reviewed the courses of crystallization of pyroxenes from mafic magmas and drew a trend line for the lime-rich clinopyroxenes. The present investigation was undertaken primarily to determine more closely the trend in the ferroaugite region, for here Hess's trend was based on the analyses of two pyroxenes from the Skaergaard intrusion and both of these are now known to be somewhat unusual types. The Skaergaard intrusion which is a strongly fractionated gabbroic complex is particularly suitable for this study of the ferroaugites. Fifteen new analyses have been made for this investigation and twenty-three analyses of the pyroxenes from this intrusion are now available. Using nineteen of these, new trend lines have been drawn (fig. 1) for the Skaergaard clinopyroxenes.

The Skaergaard intrusion of east Greenland, Tertiary in age, was described by Wager and Deer in 1939. They showed it to be a strongly differentiated layered gabbro complex having the form of an inverted cone. A uniform chilled margin to the border group gives the composition of the original magma. This rock is an olivine-gabbro of tholeiitic type (Tilley, 1951) chemically similar to the porphyritic central magma type of the British Tertiary Hebridean province. The rocks of the layered series range from hypersthene-olivine-gabbros in the lowest exposed rocks to fayalite-quartz-gabbros at the top.

Wager and Deer (1938, 1939) have published analyses of eight clinopyroxenes from this intrusion, five of which came from the layered series. They investigated the mutual relations of the clino- and orthopyroxenes in the rocks, concluding that in the earlier rocks both ortho- and clinopyroxene crystallized, but that in the middle gabbros it may have been

hat clinopyroxene alone crystallized from the magma and that some nverted to orthopyroxene with abundant clinopyroxene inclusions. In he latest stages pyroxenes with very unusual textures for igneous yroxenes occur and the authors suggested that these were really inverted orms, having crystallized from the magma as iron-wollastonites. The ew chemical data have confirmed the suggestions of Wager and Deer egarding the inversion origin of these pyroxenes.

In their discussion on the trend of the pyroxenes Wager and Deer onsidered that in the early stages two pyroxenes (diopsidic augite and rthopyroxene), crystallized out, and, with progressive fractionation of e magma, both became enriched in iron, while at the same time the inopyroxene became progressively poorer in lime. Eventually ortho-yroxene ceased to crystallize and they considered that this was due to e increasing capacity of the clinopyroxene to take the whole of the yroxene components into solid solution as the iron content increased. hus, towards the iron-rich end, the immiscibility gap between lime-ch clinopyroxenes and pigeonites narrowed and eventually, beyond f (orthoferrosilite) 65–70, only one moderately lime-poor phase crystal-zed. They termed such pyroxenes 'clinopyroxenes of pigeonite type'.

Hess (1941) pointed out that these iron-rich clinopyroxenes of the kaergaard intrusion were not true pigeonites, but rather, iron-rich embers of the augite series, i.e. ferroaugites. He considered that the eld of orthopyroxene (in this case inverted pigeonite) was limited by ne incoming and enlargement of the olivine field as the magmatic liquid ecame increasingly rich in iron.

The additional analyses carried out for this paper have revealed that ne trend of the clinopyroxenes shows two abrupt changes in direction. hese coincide with the disappearance of olivine as an orthomagmatic ineral at the top of the hypersthene-olivine-gabbros (height 900 metres the layered series) and with its later reappearance in the lower ferro-abbros (height 1400 metres in the layered series).

The early clinopyroxenes found in the hypersthene-olivine-gabbros e lime-rich diopsidic augites. In the olivine-free middle gabbros, which cceed the hypersthene-olivine-gabbros, the composition of the clino-yroxene becomes increasingly poorer in lime and richer in iron. In the wer ferrogabbros the lime content of the clinopyroxene increases again. ith increasing iron content the trend of compositions of the normal rroaugites turns towards hedenbergite with their lime content main-ining a gradual increase. The clinopyroxenes formed by the iron-ollastonite–hedenbergite inversion, however, become increasingly more

iron-rich and lime-poor over the range in which they crystallize and thus in the latest stages two distinct trends can be recognized.

In this paper the author has followed Hess (1949) in expressing the compositions of the clinopyroxenes as atomic percentages in terms of the Ca, Mg, and total Fe components. A complete description of the rocks and their textures is given by Wager and Deer (1939).

II. Discussion of analyses.

In the earlier rocks of the intrusion, where two pyroxenes occur, it was not possible to effect a separation of the two minerals and the mixed pyroxenes of the rock were therefore analysed (table I). Analyses 18

Table I. Analyses of mixed pyroxenes.

Analysis no.	2.	5.	6.	7.	8.	9.	A.	B.
SiO_2 ...	50·58	51·66	49·89	47·66	48·97	47·92	46·29	46·05
Al_2O_3 ...	4·06	2·44	3·74	3·48	3·85	4·60	2·26	1·17
TiO_2 ...	0·55	0·99	0·93	0·81	0·81	1·71	0·74	0·66
Fe_2O_3 ...	1·39	0·60	0·21	0·38	2·18	0·71	1·36	0·60
FeO ...	11·07	9·37	12·81	14·57	11·93	18·85	29·68	31·10
MnO ...	0·25	0·15	0·27	0·30	0·28	0·35	0·90	0·78
MgO ...	22·07	15·14	14·59	13·45	13·96	12·03	0·69	1·01
CaO ...	10·01	18·74	16·83	17·94	17·36	12·95	16·07	17·64
Na_2O ...	0·16	0·46	0·43	0·42	0·38	0·51	0·73	0·73
K_2O ...	0·01	0·14	0·13	0·18	0·21	0·13	0·19	0·19
H_2O+ ...	0·16	0·11	0·16	0·11	0·12	0·20	0·43	0·43
H_2O- ...	0·09	0·11	0·07	0·11	0·11	0·18	0·20	0·20
	100·40	99·91	100·06	99·41	100·16	100·14	99·54	100·56

A. Fraction consisting of 96 % green pyroxene from rock 4143.
B. Fraction consisting of 48 % brown pyroxene from rock 4143.
See foot of table II.

and 22 (table II) represent the brown and the green pyroxenes respectively occurring together in a rock at a height of 2400 metres in the layered series. The refractive indices of the two varieties are similar, but the optic axial angle is 55° for the brown variety and 52° for the green. An attempt was made to separate the two varieties and by repeated centrifuging two fractions were obtained, the heavier (analysis A, table I) consisting of 96 % of the green, and the lighter (analysis B, table I) about 48 % of the brown.

In general, there is a decrease in the silica weight percentage of the pyroxenes as the layered series is ascended, and this is due to the replacement of Mg by Fe″. Aluminium, at its maximum in the pyroxene of the chilled margin, remains fairly constant in the earlier augites, but

TABLE II. Analyses and optical properties of clinopyroxenes.

	1.	2a.	5b.	6c.	7d.	8e.
SiO_2	49·84	49·98	51·48	48·98	46·80	48·60
Al_2O_3	6·22	4·49	2·87	4·67	4·56	4·26
TiO_2	0·90	0·61	1·16	1·16	1·07	0·90
Fe_2O_3	2·86	1·53	0·70	0·25	0·50	2·41
FeO	8·67	9·81	6·26	9·89	9·89	9·85
MnO	0·23	0·28	0·18	0·33	0·39	0·31
MgO	12·67	21·76	14·35	12·72	12·29	13·57
CaO	17·80	11·07	22·04	21·02	23·43	19·20
Na_2O	0·38	0·18	0·54	0·53	0·56	0·42
K_2O	0·05	0·01	0·16	0·16	0·23	0·23
H_2O+	0·18	0·18	0·13	0·20	0·14	0·13
H_2O-	0·07	0·10	0·13	0·09	0·14	0·12
	99·87	100·00	100·00	100·00	100·00	100·00
... ...	1·695	1·672	1·690	1·695	1·700	1·698
... ...	—	1·678	1·696	1·702	1·706	1·705
... ...	1·719	1·701	1·718	1·722	1·730	1·724
$\gamma-\alpha$	0·024	0·029	0·028	0·027	0·030	0·026
$2V_\gamma$	43°	42°	42°	43°	42°	40°
$\gamma:c$	41°	39°	40°	41°	36°	40°
Sp. gr.	—	3·35	3·34	3·38	3·38	3·38
Atomic %						
Ca	40	22	46	45	49	40
Mg	38	60	42	37·5	33	40
Fe	22	18	12	17·5	18	20
$\frac{FeO+Fe_2O_3\times 100}{MgO+FeO+Fe_2O_3}$						
mineral	46	34	33	43	46	47
rock	55	49	—	55	—	—

	9f.	10.	11.	12A.	13.	14.
SiO_2	47·55	48·79	49·38	49·38	49·03	47·70
Al_2O_3	5·01	1·91	0·93	2·93	0·75	1·55
TiO_2	1·86	1·00	0·79	0·71	0·85	1·24
Fe_2O_3	0·78	2·90	2·09	1·48	3·07	1·82
FeO	17·56	18·13	18·46	17·36	18·62	20·32
MnO	0·38	0·17	0·29	0·23	0·31	0·38
MgO	11·63	11·06	10·98	10·66	10·16	8·73
CaO	14·12	15·58	16·43	16·83	16·79	17·32
Na_2O	0·55	0·43	0·59	0·37	0·47	0·65
K_2O	0·14	0·26	0·07	0·11	0·13	0·13
H_2O+	0·22	0·12	0·19	0·15	0·17	0·21
H_2O-	0·20	0·14	0·15	0·07	0·14	0·11
	100·00	100·49	100·35	100·28	100·49	100·16
α	1·707	1·704	1·704	1·709	1·712	1·718
β	1·715	1·712	1·711	1·713	—	1·725
γ	1·735	1·733	1·733	1·739	1·738	1·741

TABLE II (*cont.*)

$\gamma-\alpha$	0·028	0·029	0·029	0·030	0·026	0·023
$2V_\gamma$	40°	45°	45°	46°	52°	52°
$\gamma:c$	40°	39°	41°	44°	43°	43°
Sp. gr.	3·42	—	—	3·49	3·50	3·51
Atomic %						
Ca	31	33	34	36	35	37
Mg	36	33	32	32	29	26
Fe	33	34	34	32	35	37
$\frac{FeO+Fe_2O_3\times 100}{MgO+FeO+Fe_2O_3}$						
mineral	61	66	65	64	68	72
rock	70	—	—	79	—	86

	15.	16.	17.	18.	19.	20.
SiO_2	47·45	46·06	47·20	45·05	46·54	—
Al_2O_3	2·57	4·06	1·90	nil	5·77	—
TiO_2	1·14	1·58	1·21	0·55	1·22	—
Fe_2O_3	2·83	1·21	1·26	nil	1·88	—
FeO	21·54	24·02	27·43	31·76	24·65	—
MnO	0·45	0·42	0·20	0·64	0·74	—
MgO	6·30	4·51	2·54	1·35	0·79	—
CaO	16·83	17·37	17·78	19·09	17·70	—
Na_2O	0·64	0·53	0·56	0·73	0·62	—
K_2O	0·25	0·21	0·26	0·20	0·15	—
H_2O+	0·30	0·21	0·29	0·43	0·39	—
H_2O-	0·18	0·12	0·11	0·20	0·18	—
	100·48	100·30	100·74	100·00	100·63	—
α	1·719	1·735	1·737	1·745	1·726	1·728
β	1·726	1·741	1·741	—	1·732	1·735
γ	1·747	1·761	1·765	1·770	1·753	1·756
$\gamma-\alpha$	0·028	0·026	0·028	0·025	0·027	0·028
$2V_\gamma$	47°	51°	50°	55°	49°	56°
$\gamma:c$	47°	40°	43°	—	44°	—
Sp. gr.	—	—	—	—	3·48	—
Atomic %						
Ca	38	38	41	41	44	—
Mg	20	14	8	4	3	—
Fe	43	48	51	55	53	—
$\frac{FeO+Fe_2O_3\times 100}{MgO+FeO+Fe_2O_3}$						
mineral	79	84	92	96	97	—
rock	—	94	—	—	98	—

	21.	22.	23.	3.	4.	12.
SiO_2	45·93	46·60	42·62	50·39	49·68	42·32
Al_2O_3	2·25	2·37	5·24	3·54	4·99	2·25
TiO_2	0·71	0·75	1·69	0·87	1·17	4·42
Fe_2O_3	1·98	1·42	3·74	1·95	1·12	4·72

TABLE II (*cont.*)

FeO	28·94	29·70	31·54	8·47	11·76	25·13
MnO	1·16	0·92	0·78	0·19	0·16	0·22
MgO	0·79	0·67	0·47	15·82	12·79	8·33
CaO	15·54	16·01	12·27	18·41	17·97	12·07
Na_2O	0·78	0·73	1·02	0·70	0·56	0·61
K_2O	0·21	0·20	0·23	0·14	0·13	trace
H_2O+	0·44	0·43	0·48	0·11	0·18	0·12
H_2O-	0·22	0·20	0·22	0·04	0·15	0·25
	98·95	100·00	100·30	100·63	100·66	100·44
α	1·747	1·745	1·743	1·684	1·690	1·721
β	1·754	1·753	1·751	1·692	1·697	1·728
γ	1·772	1·771	1·772	1·712	1·716	1·749
$\gamma-\alpha$	0·025	0·026	0·029	0·028	0·026	0·028
$2V_\gamma$	50°	52°	58°	46°	42–52°	44°
$\gamma:c$	39°	41°	47°	39°	38°	39°
Sp. gr.	—	—	3·65	3·35	3·37	3·50
Atomic %						
Ca	37·5	38	30	38	39	26
Mg	2·5	2	2	45	39	32
Fe	60	59	68	17	22	49
$\frac{FeO+Fe_2O_3\times 100}{MgO+FeO+Fe_2O_3}$						
mineral	98	98	99	40	49	87
rock	99	—	96	—	—	79

KEY TO TABLES I AND II

1. Pyroxene from chilled margin. Rock no. 1724.
2. Augite from transitional layered series. Rock 4087.
3. ,, ,, lower border group. Rock 4289.
4. Pyroxene from border group. Rock 1222.
5. ,, ,, hypersthene-olivine-gabbro, height 300 metres (Rock 4084).
6. ,, ,, ,, ,, ,, ,, 500 m. (4077).
7. ,, ,, ,, ,, ,, ,, 700 m. (2307).
8. ,, ,, middle gabbro, height 1100 m. (1691).
9. ,, ,, ,, ,, ,, 1150 (3662).
10. Ferroaugite from hortonolite-ferrogabbro, height 1600 m. (2580).
11. ,, ,, ,, ,, ,, 1700 m. (2573).
12. ,, ,, ,, ,, ,, 1800 m. (1907).
12s. ,, ,, ,, ,, ,, 1800 m. (1907).
13. ,, ,, ,, ,, ,, 1900 m. (4265).
14. ,, ,, ,, ,, ,, 2000 m. (4272).
15. ,, ,, ferrohortonolite-ferrogabbro, height 2100 m. (4146).
16. ,, ,, ,, ,, ,, 2200 m. (4145).
17. ,, ,, ,, ,, ,, 2300 m. (4144).
18. Brown pyroxene from ferrohortonolite-ferrogabbro, height, 2400 m. (4143).
19. Green pyroxene from unlaminated layered series, height 2750 m. (4136).
20. ,, ,, ,, transgressive granophyre (1223).
21. ,, ,, ,, ferrohortonolite-ferrogabbro, height 2375 m. (1974).
22. ,, ,, ,, ,, ,, ,, 2400 m. (4143).
23. ,, ,, ,, fayalite-ferrogabbro ,, 2500 m. (1881).

TABLE II (*cont.*)

Note.—All refractive indices ±0·002. 2V and $\gamma:c \pm 2°$.

Pyroxene numbers 1–8 colour light brown; pleochroism weak.
,, ,, 9–18 colour brownish; pleochroism weak, α yellow-brown, β violet-brown, γ violet-brown.
,, ,, 19–20 pale greenish-brown, pleochroism weak,
,, ,, 21–23 greenish; pleochroism, α pale green, β green, γ green.

a, Clinopyroxene, analysis 2, after subtracting 9·6 % hypersthene (Of 41) and recalculating to 100 %.
b, ,, ,, 5, after subtracting 14·8 % hypersthene (Of 44) and recalculating to 100°.
c, ,, ,, 6, after subtracting 20 % hypersthene (Of 45) and recalculating to 100 %.
d, ,, ,, 7, after subtracting 23 % hypersthene (Of 49) and recalculating to 100°.
e, ,, ,, 8, after subtracting 9·8 % hypersthene (Of 50) and recalculating to 100 %.
f, ,, ,, 9, after subtracting 8·4 % hypersthene (Of 58) and recalculating to 100%.

Analyses 2, 3, 4, 6, 9, 12, 19, and 23 by W. A. Deer (Wager and Deer, 1939).
Analyses 1, 5, 7, 8, 10, 11, 12A, 13, 14, 15, 16, 17, 18, 21, and 22 by I. D. Muir.

later shows a general tendency to decrease. It might be considered that the higher alumina in the recalculated analyses could be attributed to its presence in the associated orthopyroxene, for no allowance was made for this constituent in the subtraction. In the Skaergaard rocks selected, however, orthopyroxene never exceeds one quarter of the total pyroxene, and a survey of recent reliable analyses indicated that orthopyroxenes seldom carry much in excess of 2 % Al_2O_3. It is clear, therefore, that the bulk of the alumina of the analysis must be present in the monoclinic variety. Titania shows a general tendency to increase from about 1 % in the earlier pyroxenes to 1·5 % in the later ones, but the behaviour of this constituent can be demonstrated much more clearly when considered in relation to the structure. Both manganese and total alkalis also show a general tendency to increase in the later pyroxenes, as does ferric iron.

An attempt was made to separate the pyroxene from a contaminated rock of the border group where hypersthene-olivine-gabbro is hybridized by the country gneiss. These hybrids become very coarse-grained, having pyroxene crystals up to 10 mm. in length. These, and strongly zoned plagioclase crystals, are set in a granophyric groundmass. From the evidence of drusy cavities, volatile constituents were clearly present in many of the granophyre hybrids and it seems likely that a vapour phase separated at a late stage in the crystallization.

G. C. Kennedy (1949), discussing equilibrium relations between iron oxides and volatiles in igneous rocks, has shown experimentally that the

Fe_2O_3: FeO ratio in a rock melt depends on the partial pressure of oxygen with which the melt is an equilibrium and that the reaction $2Fe_2O_3 \rightleftharpoons 4FeO + O_2$ may be expressed as a close approximation by the law of mass action. From this it follows that at any given temperature an increase in the partial pressure of oxygen should drive the reaction to the left. With falling temperature, however, and a constant partial pressure of oxygen, the reaction would tend to proceed to the right. In other words, at a lower temperature a greater increase in the partial pressure of oxygen is required to increase the Fe_2O_3: FeO ratio. No analysis of the granophyre is available, but considerable magnetite is present in the rock. It was thought that the composition of the pyroxene might give an indication of whether this increase in the Fe_2O_3: FeO ratio took place in the magma. The optical properties of the pyroxene are α 1·692, γ 1·722, 2V 52° (centre) and 49° (margins). As much of the pyroxene has been converted to a pale amphibole, it was impossible to separate sufficient pure material for an accurate analysis, but on 0·35 gram separated and with about 5 % of green amphibole remaining, the figures obtained were FeO 10·41, Fe_2O_3 7·77 %. These tend to support the views put forward by Kennedy. Another feature of the analysis of the pyroxene worthy of mention is that hybridization of hypersthene-olivine-gabbro with the gneiss produces a pyroxene in hybrid rocks whose composition, except for the Fe_2O_3: FeO ratio, is similar to those from the middle gabbros.

The most marked feature of the analyses is the complementary variation in magnesium and ferrous iron. When plotted against height in the layered series both MgO and FeO form smooth curves. The replacement of magnesium by ferrous iron in progressively later pyroxenes is now well known. The other important variable in the analyses is the lime content. In the layered series the lime content of the augites remains almost constant during the hypersthene-olivine-gabbro stage, but in the middle gabbros the clinopyroxenes become increasingly lime-poor and pass in the field of augites from olivine-free tholeiitic rocks. In the early ferrogabbros, however, the lime content increases again, and the only comparison with this known to the author is that shown by the later pyroxenes of the New Amalfi sheet described by Poldervaart (1944), where pyroxenes remarkably similar in texture to the inverted varieties occur. It will be noted that the lime content decreases again in the green pyroxenes considered to have formed by inversion from iron-wollastonite, whereas in the normal brown pyroxenes it increases until in the latest stages the composition approaches closely to hedenbergite.

III. The crystal chemistry of the pyroxenes.

Warren and Bragg (1928) first worked out the structure of diopside, and Warren and Biscoe (1931) showed that diopside, hedenbergite, acmite, jadeite, spodumene, augite, and clinoenstatite had essentially similar structures of which diopside could be taken as the type. They suggested as a general formula for members of the pyroxene group $X_m\ Y_{(2-m)}z_2(O,OH,F)_6$; where X = Ca, Na, (K), (Mn), (Mg), Y = Mg, Fe″, Mn, Fe‴, Ti, Al, (Zn), z = Si, Al.

Dixon and Kennedy (1933) showed that in some cases it was necessary to assume a replacement of silicon by titanium in order to make the analysis fit the formula. They listed a number of analyses in which combined silicon and aluminium was insufficient to supply the z group in the centre of the tetrahedron. Barth (1931) came to a similar conclusion with regard to the titaniferous pyroxene from Hiva Oa, as did Deer and Wager (1938) for pyroxene no. 12 of the Skaergaard intrusion.

The early clinopyroxenes are normal gabbroic types with more than sufficient aluminium to satisfy the silicon deficiency in the tetrahedral positions, and some must therefore be present in the Y group. This condition appears to hold to the end of the middle gabbro stage, but in the pyroxenes of the early ferrogabbros there is insufficient aluminium to satisfy the requirements of the z group, and in pyroxene no. 10 titanium is required to bring this group up to 2 (table III). A similar assumption must be made to balance the electrostatic charges for pyroxenes 11, 12, 13, and 14. In pyroxene no. 15 there is just sufficient aluminium present to satisfy the z group, and in no. 16 and all succeeding analyses aluminium is again in excess and some must be present in the Y group. It would appear, therefore, that over a limited range of the ferrogabbros titanium occupies positions in the z group of the pyroxenes.

In the analyses where no aluminium occurs in the Y group there is an increase in the content of ferric iron in the pyroxenes, and this replacement of Fe″ by Fe‴ is required to balance the substitution of Al for Si. Thus in analyses 10, 11, 12, 13, and 14, ferric iron is present in much greater amount than in the other pyroxenes, except for the inverted ones where a high ferric iron content is required partly to balance the aluminium and partly to offset the substitution of sodium and potassium for calcium in the X group, for the alkalis show a significant increase at this stage.

In the three earliest analyses from the layered series the (Ca+Na+K) comprising the X group remains fairly constant at about 0·9. At this stage, therefore, any replacements that occur must be due to the

TABLE III. Formulae of pyroxenes on basis of six oxygen atoms.

		1.	2.	5.	6.	7.	8.	9.	10.
	Si	1·857	1·836	1·911	1·822	1·793	1·840	1·833	1·889
	Al	0·143	0·165	0·089	0·178	0·207	0·160	0·167	0·086
	Ti	—	—	—	—	—	—	—	0·025
	Al	0·133	0·029	0·032	0·027	—	0·031	0·060	—
	Ti	0·025	0·017	0·033	0·032	0·032	0·025	0·054	0·006
	Fe'''	0·082	0·043	0·018	0·007	0·021	0·068	0·023	0·084
	Fe''	0·269	0·301	0·198	0·308	0·315	0·311	0·566	0·558
	Mn	0·006	0·009	0·007	0·010	0·013	0·008	0·012	0·004
	Mg	0·675	0·190	0·759	0·705	0·706	0·776	0·668	0·645
	Ca	0·717	0·435	0·877	0·837	0·961	0·777	0·583	0·649
	Na	0·023	0·013	0·044	0·038	0·041	0·027	0·033	0·036
	K	0·002	—	0·013	0·007	0·009	0·009	0·007	0·014
	z	2·00	2·00	2·00	2·00	2·00	2·00	2·00	2·00
	XY	1·94	2·04	1·98	1·97	2·10	2·03	2·01	2·00

		11.	12A.	13.	14.	15.	16.	17.	18.
	Si	1·916	1·905	1·920	1·886	1·881	1·850	1·918	1·916
	Al	0·042	0·095	0·037	0·074	0·119	0·150	0·082	—
	Ti	0·031	—	0·026	0·037	—	—	—	0·022
	Al	—	0·011	—	—	—	0·042	0·011	—
	Ti	—	0·021	—	—	0·033	0·047	0·036	—
	Fe'''	0·062	0·042	0·089	0·049	0·086	0·036	0·039	—
	Fe''	0·595	0·560	0·573	0·672	0·712	0·804	0·926	1·122
	Mn	0·009	0·007	0·010	0·012	0·014	0·014	0·007	0·022
	Mg	0·639	0·618	0·604	0·520	0·376	0·271	0·155	0·089
	Ca	0·685	0·694	0·703	0·736	0·717	0·747	0·773	0·847
	Na	0·046	0·029	0·037	0·050	0·043	0·041	0·049	0·056
	K	0·004	0·004	0·004	0·007	0·014	0·011	0·015	0·010
	z	1·99	2·00	1·98	2·00	2·00	2·00	2·00	1·94
	XY	2·04	1·99	2·02	2·05	2·00	2·01	2·01	2·14

		19.	21.	22.	23.	3.	4.	12.
	Si	1·867	1·928	1·925	1·785	1·836	1·862	1·720
	Al	0·133	0·072	0·075	0·215	0·152	0·138	0·107
	Ti	—	—	—	—	—	—	0·135
	Al	0·145	0·039	0·045	0·043	—	0·083	—
	Ti	0·037	0·022	0·025	0·053	0·024	0·033	—
	Fe'''	0·057	0·062	0·045	0·118	0·053	0·039	0·144
	Fe''	0·827	1·012	1·030	1·105	0·258	0·369	0·853
	Mn	0·025	0·041	0·035	0·028	0·006	0·005	0·008
	Mg	0·048	0·050	0·042	0·029	0·859	0·714	0·503
	Ca	0·761	0·698	0·709	0·550	0·719	0·721	0·525
	Na	0·048	0·063	0·055	0·083	0·049	0·041	0·048
	K	0·008	0·011	0·010	0·012	0·008	0·006	—
	z	2·00	2·00	2·00	2·00	1·99	2·00	1·96
	XY	1·96	2·00	1·99	2·02	1·99	2·01	2·08

substitution of Mg by Fe″. In the middle gabbros, however, the X group decreases sharply to 0·62 in pyroxene no. 9, and here the iron introduced must be replacing the calcium, for magnesium remains almost constant. In the lower ferrogabbros (analyses 10–14) the X group increases again, while magnesium decreases, its place being taken by ferrous iron. This process continues with iron replacing magnesium and with a gradual increase in the X group until the latest stages in the normal pyroxenes.

The changes in the trend of the pyroxenes can be understood when the structure is considered, and this matter has been discussed by Edwards (1942, p. 598) and by Walker and Poldervaart (1949, p. 643), who have pointed out that the entry of the small Fe″ ion in place of the larger Ca″ ion (seen in the middle gabbro pyroxenes) would cause a considerable distortion of the lattice, and that such a replacement would have a limiting value beyond which internal stresses would render the structure unstable. This position appears to be almost reached in pyroxene no. 9, and further enrichment in iron should now take place only at the expense of magnesia. The ferrous ion is slightly larger than the magnesium ion, and thus, as more iron is introduced, the interatomic distances between the chains would be slightly increased. This would further restrict the limits of the iron replacement of calcium, and thus an increasing proportion of calcium ions would again tend to enter the pyroxene structure as the ferrohedenbergite end of the series is approached.

In plutonic rocks, with slow cooling, it may be that the clinopyroxene structure only remains stable when the replacement of calcium by ferrous iron is less than about one-third, and any further replacement by iron must therefore come at the expense of magnesium. The only clinopyroxenes known in which the replacement of calcium by ferrous iron is greater than about a third are the pigeonites and the slag pyroxene described by Bowen (1935). Many of these are the products of rapid crystallization and are probably metastable forms.

IV. The change in composition of the pyroxenes and their relations to the magma and the other mineral phases.

The clinopyroxenes of the layered series form an unbroken series, and trend lines have been drawn from them based on 19 analyses (fig. 1). Wager and Deer (1939), from the known thermal data of the systems clinoenstatite–diopside, hedenbergite clinoferrosilite, and clinoenstatite–clinoferrosilite, have considered the probable form of the isotherma

rface within the four-component system, and this idea has been further plified by Hess (1941) who drew a series of hypothetical sections ross the area between the Di Hd and En–Fs joins to show the relaons of the lime-rich clinopyroxenes to the pigeonites. These relations

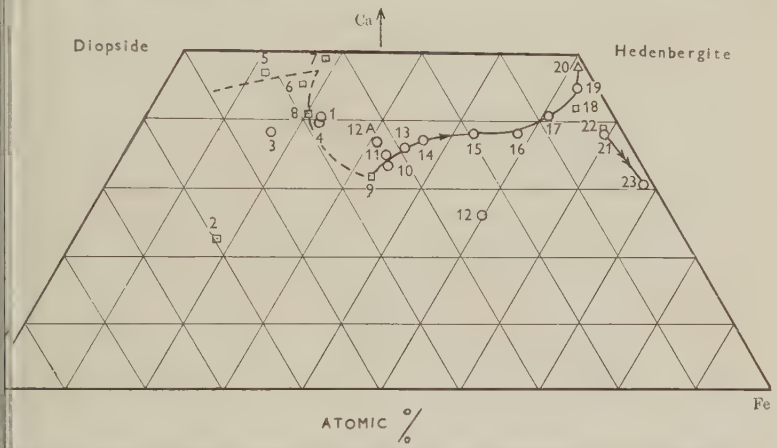

Fig. 1. Trend of clinopyroxenes of the Skaergaard intrusion, east Greenland.

○ Chemical analysis of clinopyroxene.

□ Clinopyroxene composition recalculated from analysis of bulk pyroxene of the rock.

△ Composition estimated from optical determinations.

→ Trend of normal pyroxenes, →→ trend of inverted pyroxenes.

licate a cotectic curve beginning close to the clinoenstatite diopside n near Wo 45, En 55, and extending towards the hedenbergite–moferrosilite join at Wo 30, Fs 70. As the cotectic curve approaches e two-pyroxene boundary, the solidus and liquidus of the ferroaugites nd slightly downwards and develop into a solid solution series with minimum, the eutectic between ferroaugite and ferropigeonite becomg a reaction point. From this stage onwards, the field of olivine rrows down the pigeonite field until the latter disappears altogether. om these considerations it is clear that for any position on the liquidus e corresponding solidus point should be nearer to the clinoenstatite–opside join, and hence the pyroxenes should always show enrichment magnesia compared with the liquid from which they have crystallized. is is always the case with the Skaergaard pyroxenes.

Because of the strong fractionation of the Skaergaard intrusion, the mpositions of the successive liquids and of the pyroxenes with which y are in equilibrium should give the positions of the liquidus and

solidus of the system from near the diopside–clinoenstatite join in th early rocks, to the hedenbergite-clinoferrosilite join in the latest stages The resemblance between the solidus as determined from the Skaergaar pyroxenes and that deduced from data of artificial systems (Hess, 1941 should be only approximate however, because of the extremely gentl slope of the solidus surface on the lime-rich side of the cotectic curve Because of this gentle slope of the solidus surface towards the Ca com ponent of the system it is very probable that the position of this cotecti curve and the lime content of the lime-rich solidus phase would be very susceptible to change by:

(*a*) the addition of Al‴, Fe‴, Ti‴, or similar ions to the solidus phase in any considerable amount;

(*b*) structural strains in the pyroxene caused by the replacements o Mg″ and Ca″ by Fe″ (discussed in the previous section);

(*c*) the relative amounts of the total Ca, Mg, and Fe from the magma available for incorporation in the pyroxenes (see later).

In a simple 3-component system it is this last factor which determines the position of the cotectic curve; but in a magma, phases such as plagio clase, apatite, olivine, and iron ores are competing for the available Ca, Mg, and Fe from the liquid. All these do not appear in the artificia metasilicate system $CaSiO_3$ $MgSiO_3$ $FeSiO_3$ for which the position o the cotectic curve and slope of the isothermal surface have been deduced

As Al‴, Fe‴, Ti⁗, or other ions do not enter the Skaergaard pyro xenes appreciably their effect is likely to be small and so the final position of the cotectic curve should depend mainly on the other two factors The first of these having been discussed already, attention will now be focused on the other. The crystallization of the clinopyroxenes can be resolved into the variation of three principal components—lime, magnesia and ferrous iron. As all three are available in the magma in varying amounts throughout the crystallization history, one of the controlling conditions will be the nature of the phases separating at any one time for the oxides with which the magma is saturated must be shared amongst these.

The lime taken out of the magma enters plagioclase felspar, mono clinic pyroxene, and apatite. The MgO and FeO are required to form olivine, ortho- and clinopyroxenes, and the iron ores. If the composition of the liquid varies in such a way that it moves out of the field where all these minerals can crystallize in equilibrium with each other, then one phase at least will cease to crystallize. The magma, however, would continue to remain saturated with the components which went to make

up that phase, and with falling temperature these components should be taken up by the remaining minerals which are still being precipitated. Thus, the proportions of the various oxides now entering into these minerals would be altered, and changes in their relative amounts and in composition could be expected in the other minerals which continue to separate when one mineral ceases to crystallize or a new one begins to form.

Wager and Deer (1939, p. 127) considered that the material forming the rocks of the layered series consisted of two parts: (*a*) the primary precipitate, made up of crystals which separated from the overlying magma; and (*b*) the interprecipitate material, which crystallized from the magma trapped between the crystals of the primary precipitate. Only one new mineral phase involving lime appears; this is apatite which does not occur in the primary precipitate until a late stage when the P_2O_5 content of the magma has increased to saturation point. The plagioclase felspar requires less lime as the differentiation proceeds, for it changes in composition from An 60 in the early rocks to An 30 in the latest. The lime content of the liquid decreases from about 11 % in the original magma to 6 % when the latest rocks of the layered series were formed. Hence, there is never any great impoverishment in this constituent in the later stages of the Skaergaard magma.

The ferromagnesian minerals present a somewhat more complicated story. Olivine begins to crystallize in the early rocks of the layered series, but ceases to form during the middle gabbro stage. Its reappearance marks the beginning of the ferrogabbros and it remains as an essential constituent of the rocks until the latest stages when its composition is Fa 98. Orthopyroxene crystallizes in the earliest rocks along with the clinopyroxene and continues to form until some time after the beginning of the lower ferrogabbros. It occurs in the lowest rocks of the layered series, only as an interprecipitate mineral, and here the textures do not suggest that it has inverted from a pigeonite. In the middle gabbros, however, it is intimately associated with the augite and here it is clear that both pyroxenes must have crystallized at the same stage. The orthopyroxene of such rocks contains numerous bleb-like inclusions of lime-rich clinopyroxene and would appear to have inverted from a pigeonite. When the percentage of $FeSiO_3$ in the orthopyroxenes is plotted against height, and the curve is extrapolated to the height where the mineral disappears, the latest orthopyroxene to form has a composition near Of 70.

Iron ore is present in the early layered rocks only as an interprecipitate

mineral, but it makes its appearance as a mineral of primary precipitation about the end of the hypersthene-olivine-gabbros. Magnetite and ilmenite, the two ores involved, appear to be governed by the amount of Fe_2O_3 and TiO_2, respectively, available from the magma. Since both these oxides are subtracted from the magma in only small amount in the early rocks, they become concentrated in the later liquids, as does P_2O_5. These two phases appear suddenly in the primary precipitate and take up a certain amount of the available FeO.

The chilled margin pyroxene no. 1 is only slightly poorer in lime than the three analysed pyroxenes from the hypersthene-olivine-gabbros and has a higher iron-magnesia ratio. All the analyses of augites from the hypersthene-olivine-gabbros fall into the well-known field of clinopyroxenes from olivine-gabbros and related rocks, and analyses of similar pyroxenes are abundant in the literature. The lowest part of the main-layered series is not exposed, but the pyroxenes from this part of the intrusion must certainly be more magnesian and less iron-rich than pyroxene no. 5. If the earliest rocks of the border group can be considered as analogous with the hidden layered rocks, then these too should contain very magnesian pyroxenes. Such minerals are found in the gabbro-picrite (augite γ' 1·698, 2V 58°, therefore the composition is approx. Ca 47, Mg 48, Fe 5 atomic %; hypersthene $2V_\alpha$ 82°, and therefore the composition is Of 20) and in the 'perpendicular felspar rock' (augite γ 1·702). It would appear, therefore, to be a reasonable assumption that the composition of the pyroxenes of the hidden layered series would lie on the extrapolation of the trend line drawn through analyses 7, 6, and 5 towards the diopside–clinoenstatite join. Such compositions would conform with those of pyroxenes from the Stillwater complex of Montana (Hess, 1949) and of pyroxenes from other very magnesian rocks.

Bowen (1914) who investigated the system clinoenstatite–diopside showed that in the artificial system there was complete solid solution between the end-members. In most igneous rocks, however, two pyroxene phases are normally present and Hess (1941) has suggested that the two-pyroxene field might extend at least to a line close to the diopside clinoenstatite join before changing over to a series of solid solutions with a minimum. In this connexion the composition of pyroxene no. 2 is of interest. It is a recalculated analysis, and Wager and Deer (1939) suggested that the analysed material might have contained more orthopyroxene than was allowed for and doubted whether the estimated composition of the clinopyroxene was reliable. However, even if there was three times as much orthopyroxene in the analysed

sample as was estimated, its composition would be approximately Ca 30, Mg 60, Fe 10; i.e. the clinopyroxene remains deficient in lime and lies outside the field normal to pyroxenes from such olivine-gabbros. This analysis, therefore, may be regarded as evidence that for very magnesia-rich and iron-poor pyroxenes complete solid solution between pigeonite and diopsidic augites exists. An alternative explanation of the complete miscibility observed between clinoenstatite and diopside in the artificial system and the conflicting evidence in basic magmas was put forward by Tsuboi (1932), who suggested that at lower temperatures diopside and pigeonite would show only limited miscibility, at higher temperatures more miscibility would be shown and that at dry-melt temperatures complete solid solution would be possible. Bowen, Schairer, and Posnjak (1933) discussed this matter further.

At a height of 900 metres in the layered series, i.e. about 20 metres above the level of analysis no. 7, olivine ceases to crystallize as a primary constituent. During the middle gabbro stage, as analyses nos. 8 and 9 show, the pyroxenes become progressively more lime-poor, and while their magnesia contents remain more or less constant or even increase slightly, there is an increase in ferrous iron at the expense of lime. At this stage the composition of the magma has moved from the field where olivine can crystallize together with clinopyroxene into the clinopyroxene field itself, but nevertheless it remains saturated with magnesia, ferrous iron, and silica. The only phases continuing to crystallize which could incorporate all three components are pyroxenes. To some extent the appearance of magnetite and ilmenite in the middle gabbros must reduce the amount of ferrous iron available for the pyroxenes and the presence of the iron ores may in part explain the slight reduction in the iron content of the clinopyroxenes at this stage as shown by analysis no. 8.

If more clinopyroxene of similar lime content were to crystallize in the middle gabbros to incorporate the iron and magnesia now available, the lime content of the magma would be greatly reduced or else the plagioclase to crystallize would have to become more sodic to allow for the additional withdrawal of lime to form pyroxene. If, to compensate for this withdrawal of additional lime to form pyroxene, the amount of plagioclase to crystallize were to be reduced, this in turn would effect the soda, alumina, and silica components of the magma. It is possible, therefore, that the available magnesia and ferrous iron would be taken up by an increase in the amount of clinopyroxene to crystallize, an increase which would at the same time require only the same total

amount of lime from the magma. Therefore the clinopyroxene should become poorer in lime.

Another possible way in which the available iron and magnesia could be taken up would be by an increase in the amount of orthopyroxene to crystallize (as pigeonite). However, during the middle gabbro period the amount of this mineral to crystallize decreases gradually, although it now seems to form in the primary precipitate along with augite and not merely as an interprecipitate mineral as in the lower rocks. Furthermore, when olivine reappears in the ferrogabbros, hypersthene does not disappear at once, but is still found in rocks 150 metres above the level of the reappearance of olivine. These observations do not suggest that there is an intimate relationship between the iron-rich orthopyroxene (or pigeonite) and olivine, a conclusion which is supported by the results obtained by Bowen and Schairer (1935) for the artificial system MgO–FeO SiO_2, where the reaction relation between olivine and clinohypersthene disappears for clinopyroxenes containing more than about 25 mol. % $FeSiO_3$. In the first analysed ferroaugite after olivine reappears (no. 10) a great change in the direction of the trend is noted however.

Pyroxene no. 10 comes from a rock at a height of 1700 metres in the layered series and olivine makes its appearance some 200 metres lower down. While magnesia once again begins to fall slowly, with increasing height, iron continues to increase, but the most significant change is in the lime content. It can be explained by the incoming of olivine again in significant amount and this would require a large amount of the iron and magnesia available from the magma.

A new analysis has been made of the pyroxene from the type rock of the lower ferrogabbros (1907, Wager and Deer, 1939), for this pyroxene (no. 12) lies away from the new trend line. The new analysis (no. 12A) lies near the trend; the material for this came from the Cambridge specimen. It will be noted that the optical properties of this mineral differ from those quoted for no. 12 (table 2). In order to establish whether the optics of the pyroxene of other collections of this rock agreed with those of the Cambridge specimen, γ' (on cleavage flakes) was determined for the pyroxene from three specimens of the rock from this locality kindly submitted by Professor Wager. It was also determined for the pyroxene from a neighbouring rock (1493). The refractive index γ' for all these specimens agrees within $\pm 0{\cdot}003$ with that determined for the analysed specimen no. 12A. It was also decided to check the composition of the olivine in the Cambridge specimen by measurement of the

fractive index; this gave a value of 1·829, corresponding with 74 olecular per cent. favalite, whereas Wager and Deer give the composi- on based both on chemical analysis and on optical data, as Fa 59. his difficulty has not been resolved.

The next change of direction in the trend of the pyroxenes is seen yond the level of pyroxene no. 13, where the increase in lime content adually ceases, and the progressive enrichment in iron, with a comple- entary decrease in magnesia, is even more marked. This causes a adual flattening of the curve, and this point marks the entry of apatite a primary phase. The incoming of this mineral seems to depend on e P_2O_5 content of the magma. As no apatite crystallizes in the early ges in the primary precipitate, the P_5O_5 content of the liquid increases til the magma is saturated at about 1·5 %. This new phase, which nes in very suddenly, amounts to 7 % in the parent rock of pyroxene . 14 and so requires considerable amount of lime. As the total lime ailable in the liquid continues to decrease, less is available for the roxene.

It will be noted that with the incoming of olivine the pyroxene begins decrease in amount, a decrease which is partly offset by the increase the lime content of the pyroxene. During this stage, however, the spar increases slightly in amount but decreases in its lime content. In is way the lime available from the magma is distributed. At 1850 tres, where apatite makes its appearance, the plagioclase decreases d this too makes lime available for apatite. The pyroxene now begins increase again, as does olivine, and this is due to the great increase the iron content of the liquid and to the rapidly decreasing solubility magnesia in it. This trend in the pyroxenes continues through alyses nos. 15, 16, 17, and 18. The lime content remains about Ca 39 th a slight rising tendency, while Mg decreases from 19 to 8 and at e same time Fe increases from 43 to 55. Apatite and plagioclase felspar ntinue to crystallize until after the end of the layered series, although th are decreasing in amount and therefore more and more of the avail- le lime is free for incorporation in the pyroxene.

The brown pyroxene no. 18 shows that for the normal pyroxenes e trend continues with a gradually rising lime content towards the denbergite clinoferrosilite join, and that in the latest stages where the n content of the magma is decreasing it turns towards hedenbergite. is is shown by the analysis of the pyroxene no. 19 (from a height of 0 metres) in the unlaminated layered series. The latest stages of the aergaard intrusion are represented by hedenbergite-granophyres with

green pyroxenes whose composition must lie very close to hedenbergite itself. The optical properties determined for the pyroxene no. 20 from one of these rocks were α 1·728, β 1·735, γ 1·756, 2V 56°. This mineral is associated with an olivine near fayalite.

In the rocks at a height of about 2350 metres in the layered series the clinopyroxenes develop a very unusual habit for igneous pyroxenes, occurring as patches of interlocking disoriented grains of a green colour instead of the usual brown. These pyroxenes are generally crowded with tiny granular opaque inclusions, and Wager and Deer suggested that they were really inverted forms, having crystallized from the magma as iron-wollastonites. Associated with some of these green pyroxenes, and forming a discontinuous border to them or small internal patches in them, is a brown variety (no. 18) in crystallographic continuity with the green. This pyroxene appears to have formed when the magma temperature was lower than the minimum for the iron-wollastonite–hedenbergite inversion and to have crystallized directly from the magma as a normal pyroxene. The inversion must, therefore, have occurred before crystallization was complete.

The pyroxenes which were thought to have inverted from iron-wollastonites show a very marked reduction in the lime content and contain very small amounts of magnesia. The green pyroxenes from nos. 21 and 22 are very similar and contain only about 0·6 % MgO whereas the brown pyroxene at this stage (no. 18) carries 1·35 %. Further, although the total iron content in the magma is now beginning to decline, the iron content of these pyroxenes continues to increase, rising from Fe 59 in no. 21 to Fe 68 in no. 23.

Thus, the iron-wollastonites over the range in which they crystallize become progressively more iron rich and lime poor. This is in harmony with the results obtained by Bowen, Schairer, and Posnjak (1933) in the artificial system CaO-FeO-SiO_2. Although for iron-rich wollastonites the composition of the mineral cannot be deduced from the simple binary system, the composition of analysis no. 23 corresponds very closely with the point where, with lowest temperature in the system, olivine and iron-rich wollastonite are in equilibrium. This very low content of magnesia in the green pyroxenes at the top of the layered series accounts for the very low magnesia content in the parent rocks at this stage. Analyses of later rocks in the intrusion after the whole of the layered series had ceased to crystallize show that such rocks are somewhat richer in magnesia than the rocks at the top of the layered series. It is now clear that this is due to the higher MgO content of the normal ferroaugite

compared with the iron-wollastonites. A rather unusual feature of the analyses of the inverted pyroxenes is their notable alumina content. Analyses of minerals of the wollastonite group are characterized by very low alumina values, but in the case of these inverted pyroxenes the wollastonites contained notable amounts of ferric iron and the presence of aluminium in the z group is necessary to render the structure electrostatically neutral.

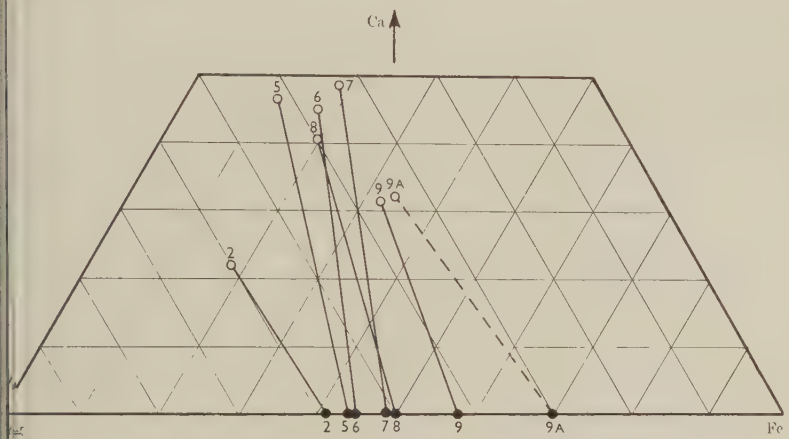

FIG. 2. Relations between clinopyroxenes and their associated orthopyroxenes. Dashed line indicates the limit of the two-pyroxene field. Atomic %.

It is of interest to compare the degree of progressive enrichment of the clinopyroxene in total iron with respect to magnesia, with that of the parent rock and with the similar enrichment in the associated orthopyroxene and olivine. For this the ratio $(FeO+Fe_2O_3+100)/(FeO+Fe_2O_3+MgO)$ is used (table II). The comparison with the parent rock shows that in all cases the ratio is lower in the pyroxene than it is for the rock, but that with gradual enrichment in iron the difference becomes less and less marked, and in the final stages the ratios are the same.

The relations of the clino- and orthopyroxenes from the same rock are shown in fig. 2. For this diagram the lime content of orthopyroxenes nos. 8 and 9 which are considered to have formed by inversion from pigeonite is not taken into account. From this it can be seen that the orthopyroxene is always richer in ferrous iron with respect to magnesia than the associated clinopyroxene. Even if the Ca component is ignored and the amounts of Mg and Fe are worked out as ratios, then the same enrichment of the orthopyroxene in ferrous iron with respect to

magnesia is seen to hold. This is in keeping with the relations of the two minerals, for the clinopyroxene occurs as a mineral of primary crystallization, whereas the orthopyroxene usually occurs interstitially and crystallized later. As the orthopyroxene had a composition of Of 70 when it finally ceased to crystallize, by interpolation along the trend line, the composition of the monoclinic pyroxene no. 9A at this stage

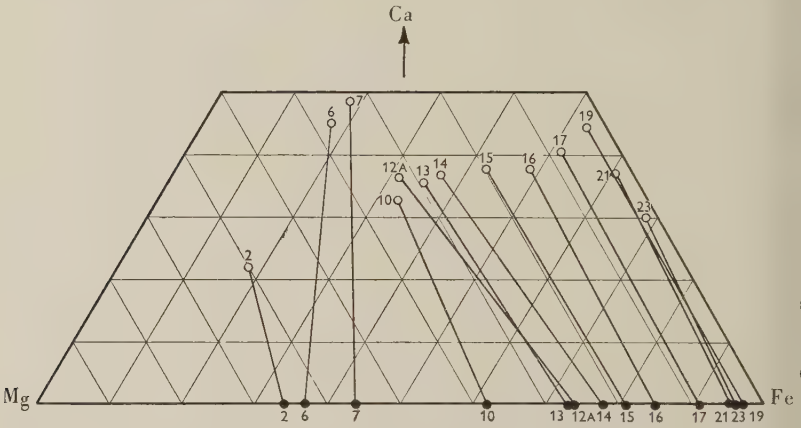

Fig. 3. Relations between clinopyroxenes and their associated olivines. Atomic %.

can be deduced. This is indicated in fig. 2 where the dashed line indicates the limit of the two-pyroxene field. The relations of the two pyroxenes, if the Ca component is ignored and the clinopyroxene is recalculated in terms of Mg and Fe, are shown in table IV.

Table IV. Comparison of iron enrichment in associated clino- and orthopyroxenes. Atomic %.

Clinopyroxene (Mg and Fe recal. to 100 %) Fe	23	27	31	35	33	48	50	51
Orthopyroxene ... Fe	41	44	45	49	50	58	70	none
No.	2	5	6	7	8	9	9A	10

In the case of the olivines, the relation between the clinopyroxenes and their associated olivines is set out diagrammatically in fig. 3. The olivines lag behind the orthopyroxenes in the replacement of Mg by Fe, a relation that is in harmony with the experimental results of Bowen and Schairer (1935). The clinopyroxenes, however, lag behind the olivines in iron enrichment in the early stages, and, with the exception of the reaction-rim olivine associated with pyroxene no. 9 which is more

agnesian than the pyroxene, the olivine remains the more iron rich until we come to the parent rocks of pyroxenes nos. 21 and 23. Here, e explanation for the reversal of the normal relationship is that iron-ollastonite was the actual phase crystallizing from the magma along th the olivine, and this, as we have seen before, was extremely poor magnesia. The relations between the clinopyroxenes and their asso-ated olivines are shown in table V.

TABLE V. Comparison of iron enrichment in associated clinopyroxenes and olivines. Atomic %.

inopyroxene ...	Fe	23	31	35	51	50	55	59	68	80	86	95	96	97
vine	Fe	34	37	44	62	74	73	78	81	85	91	97	96	96
... ...		2	6	7	10	12a	13	14	15	16	17	19	21	23

Thus, as we have seen in the Skaergaard intrusion, the orthopyroxene the most iron-rich ferromagnesian silicate, then comes the olivine, d lastly the clinopyroxene. In general, in other intrusions where the lations between the ferromagnesian minerals have been worked out on enrichment in the lime-rich clinopyroxene lags behind that of its sociated orthopyroxene or pigeonite.

V. Conclusions.

It has been shown that data from the artificial systems has enabled picture to be drawn of solidus liquidus relations in the pyroxenes. The ccursion of the olivine field limits the composition of the pigeonite-roxene series towards the iron-rich end. The extremely gentle slope the solidus surface on the lime-rich side of the cotectic curve renders e lime content of the clinopyroxene extremely sensitive to alteration y the following factors:

(*a*) the amount of Al, Fe‴, and Ti taken into solid solution;

(*b*) internal strains in the mineral structure caused by replacement of Ca and Mg by Fe″;

(*c*) by the proportions of Ca, Mg, and Fe″ available from the magma to form the clinopyroxene. In the case of the Skaergaard clinopyroxenes this last factor is considered to be the most important.

The two abrupt changes in the direction of the trend in the clino-roxene are attributed to the disappearance of olivine as a primary nase and its later sudden reappearance in the lower ferrogabbro stage. he flattening out of the trend line in the middle of the ferrogabbros ay be correlated with crystallization of apatite in large amount as a rimary mineral.

In the latest stages two distinct trends can be made out. That of th normal ferroaugites finally turns towards hedenbergite as the iron con tent of the magma begins to decline and the pyroxenes are green i colour at this stage. The other trend is that of the pyroxenes derive from the iron-wollastonite–hedenbergite inversion. Here the pyroxene become progressively more iron rich and lime poor and magnesia i reduced to insignificant amounts. The final pyroxene has a composition Ca 30, Fe 70, which corresponds to the position deduced for the lowest temperature relations in the hedenbergite–clinoferrosilite series of th artificial system.

A discussion of the significance of the Skaergaard clinopyroxene trend and comparisons with the pyroxenes from other tholeiitic rocks will b made in a later communication.

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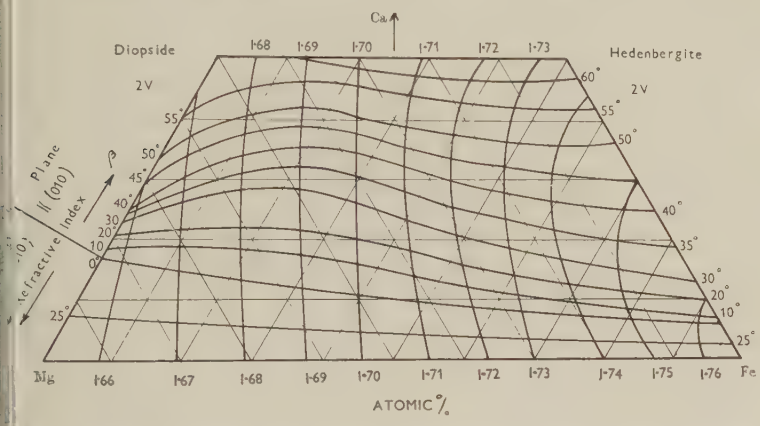
Appendix.

Note on the relationship of the optical properties to the chemical composition in clinopyroxenes.

Many attempts have been made to correlate the optical properties o clinopyroxenes with their chemical compositions and Tomita (1934) firs drew up a set of diagrams showing variations in refractive indices, opti axial angle, and extinction angles. Later these were modified by Dee and Wager (1938). Hess (1949), in a paper which lists many new chemica analyses together with very accurate optical determinations, has draw up new tables showing the variation in refractive indices, optical axia angle, and the birefringence for common clinopyroxenes. He considere that his diagrams were most reliable for those areas where natura pyroxenes are commonest and that natural pyroxenes were rare o did not exist where his data are poor. The exception was the ferro augite field where Hess's curves were based on only seven analyses, tw of which contain significant amounts of soda and so should be rejecte for 2V curves. For the present paper many new analyses of ferroaugite have been carried out and new refractive index and 2V curves hav been drawn. These should improve on the accuracy of Hess's curve for the ferroaugite field, but elsewhere they agree very closely with hi diagram. The writer has followed Hess in using atomic % Ca, Mg, F

express the compositions of pyroxenes and has found the recognition l use of the (100) parting tablets, advocated in the 1949 paper, the st convenient and rapid method of determining a refractive index of linopyroxene. The new curves are presented in fig. 4.



IG. 4. Relation between the chemical composition of clinopyroxenes and the es of 2V and the refractive index along the *b* crystallographic axis.

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Sedimentary inclusions in the hypersthene-gabbro, Ardnamurchan, Argyllshire.

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Introduction.

THE hypersthene-gabbro of Ardnamurchan Point is the earliest of the intrusions of 'Centre 2' of the Ardnamurchan complex as shown by J. E. Richey (11, p. 208).[1] The mass is intruded into domed Jurassic sediments and Tertiary basalt flows (fig. 1). The main part of the intrusion is composed of rather fine-grained gabbro, containing 2–3 % of hypersthene which typically surrounds crystals of olivine. Marginally, however, the gabbro is a quartz-bearing variety, with a micrographic mesostasis. The outer contact is somewhat sinuous due to the fact that the intrusion has only been exposed to a comparatively small depth below its original roof. This same geological accident may account in part for the number of fine-grained, basic, granular masses that are locally abundant within the gabbro. These masses, variously described as pyroxene-granulite or pyroxene-granulite-hornfels, have a granoblastic texture and are composed of normal gabbroic minerals: plagioclase, augite, olivine, hypersthene, and magnetite. They have been described in detail by J. E. Richey and H. H. Thomas (11, pp. 229–235) who considered them to be derived in part from the recrystallization of baked inclusions of basalt, and in part from cognate masses of early emplaced gabbro, prized off the walls and incorporated in the later pulses of the gabbro magma. The only other description of inclusions in the hypersthene-gabbro is that of A. G. MacGregor (6) who compared these pyroxene-granulite-hornfelses with similar rocks of other areas and concluded that they represent basaltic and gabbroic rocks that have suffered thermal, and essentially non-additive, metamorphism. The only previous mention of sedimentary rocks in the gabbro is in connexion with the sapphire-spinel-hornfels that occurs at the northern end of Glebe Hill, near Kilchoan (locality 5, fig. 1), and which Thomas believed

[1] For list of references see p. 735.

to be derived from an aluminous bole produced by weathering of Tertiary lavas (11, pp. 233–235).

The published accounts lead one to suppose, therefore, that sedimentary inclusions are virtually absent from the hypersthene-gabbro.

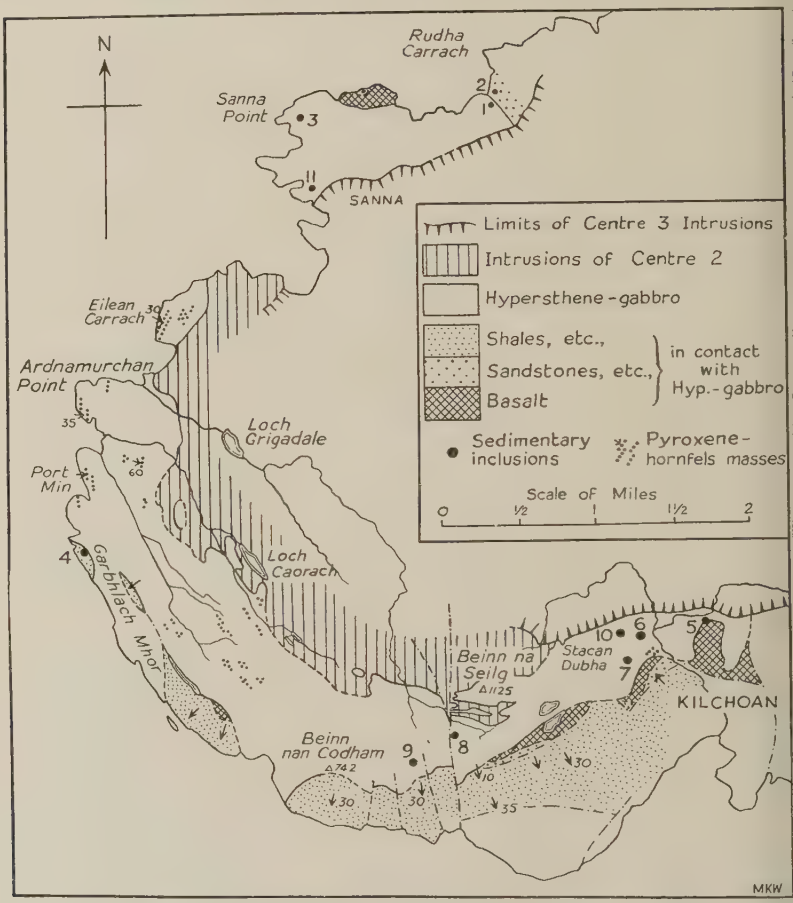

Fig. 1. Map showing localities of xenoliths in the hypersthene-gabbro. (The main boundaries have been taken from the coloured map of the Ardnamurchan memoir, 1930.)

In actual fact, xenoliths that one would class unhesitatingly as of sedimentary origin are exceedingly rare, and the writer has discovered only three occurrences (or four with the sapphire-bearing rocks already known); but there are, in addition, a number of banded granulite-

hornfelses whose origin may be considered doubtful, but which are most probably derived from sediments. They have been chemically reconstituted, and evidence regarding their origin rests largely upon their microtexture, and upon structures which are believed to be inherited. Even if this view is taken, the number of probable sedimentary inclusions is still greatly outweighed by that of inclusions of igneous appearance, and the dominance of the latter, is an intrusion emplaced largely amongst sediments, remains a major problem.

The writer proposes to describe first the few xenoliths whose sedimentary origin is not in doubt, and to follow this by a discussion of the origin of the more abundant banded or layered basic hornfelses which he believes to be sedimentogenous also. The xenoliths in the first category are in all cases surrounded by marked reaction zones. In the order in which they are described these xenoliths are: (1) siliceous rocks which have been felspathized to produce granophyre; (2) finely interbedded shaly and ferruginous rocks which have been converted to rocks of noritic composition; and (3) peraluminous sediments (bole) which have been altered to basic plagioclase-magnetite-spinel-(corundum)-rocks.

Siliceous xenoliths and their reaction with gabbro magma.[1]

The hypersthene-gabbro intrusion terminates with a steep junction against the Scalpa beds (Middle Lias) at the north-western end of its outcrop. These beds are chiefly fine-grained sandstones with subordinate shaly and calcareous layers. They are cut by numerous cone-sheets and a few small dikes, the whole assemblage being metamorphosed by the hypersthene-gabbro. Two rounded inclusions of profoundly altered sandstone, both about a foot in diameter, occur about 15 feet inside the gabbro just above high-water mark (loc. 1, fig. 1). Unfortunately this otherwise ideal exposure is marred by a black oily coating close to the rock and some of the finer points of the geology may easily be overlooked. These two xenoliths are conspicuous, however, for they have been accessed by weathering and have developed a fretted surface. Each inclusion has two parts; a very fine-textured core of light rock, speckled with coloured minerals sparsely scattered through it, and a complex marginal zone showing small-scale banding and dense concentrations of

[1] Since a number of the effects described in this paper are metasomatic, it is desirable to state that the gabbro as a whole shows evidence of a magmatic origin. This evidence will be given in a forthcoming account of the petrology and structure of the hypersthene-gabbro and adjacent intrusions.

coloured minerals (fig. 4). The mineral content and texture changes rapidly within the marginal zone so that it is possible to distinguish an inner and an outer part as described below, despite the fact that the total width is only about an inch.

The cores of the xenoliths consist dominantly of micrographic intergrowths of quartz and alkali-felspar, with aggregated crystals of green clinopyroxene. The latter is a pale mineral with the following optical properties: β 1·722±0·002 2V 63°±1°, $\gamma:c = 55°$. The pleochroism scheme is α yellowish-green, β yellow-brown, and γ bluish-green. These properties correspond most closely with those which Winchell (15, p. 191) gives for a pyroxene of the diopside-hedenbergite series with a small content of acmite, approximately $Di_{40}Hd_{50}Ac_{10}$. The crystals are ragged in outline and have a maximum diameter of about 1 mm. Plagioclase is subordinate to the micrographic alkali-felspar, and occurs as isolated well-formed grains showing quite strong zoning. The more calcic parts of these crystals are clouded with minute iron-ore inclusions in a manner that is characteristic of the plagioclase of all the surrounding gabbro. This is probably a thermal metamorphic effect due to the nearby great eucrite intrusion of 'Centre 3'. Sphene and apatite are the only other important constituents. Some of the sphene crystals are perfectly euhedral but others are completely shapeless. The apatite occurs as euhedral bluish prismatic crystals up to about 0·25 mm. in length, having clouded cores. Occasional patches of calcite can be observed, interstitial to well-formed quartz.

In one specimen, the granophyric core of the xenolith passes abruptly into the complex marginal zone, which is in two parts. The inner part consists of spongy porphyroblasts of the green pyroxene and sphene set in a groundmass of minutely granular plagioclase. The latter occurs as equant grains of almost square cross-section, which are feebly twinned and marginally zoned. Accurate refractive index measurements on the central and marginal parts of such minute crystals are difficult to make, but the zoning appears to be normal, involving a small range of composition, with a maximum anorthite content of 40. The coloured minerals are concentrated into well-marked bands. In view of the weathered-in nature of the xenoliths, it is impossible to say whether this development of unusually large quantities of sphene and of plagioclase is present all round both xenoliths, or whether it is a purely local development, dependent upon the presence of more lime than usual in one part of the original rock. Although the calcium may have been derived in part from the parent rock, the titanium cannot be accounted for in this

ny, and its presence affords a good example of a geochemical culmination of a single element which is not otherwise abundant, either in the bbro or in the xenolith rock.

The remaining outer part of the marginal zone of both xenoliths consists solely of the green pyroxene. A similar narrow zone of pyroxenite also found at the main contact which the gabbro makes with the iliments. As one passes across the pyroxenite away from the xenoliths the main mass of sediments and into the gabbro the pyroxene loses colour, and half-an-inch from the margins of the xenoliths all that nains is a feeble marginal green zoning around the normal colourless bbroic augite. Beyond that the hypersthene-gabbro assumes those aracteristics which are found with little variation in all the marginal ies of the intrusion. The most notable feature is the occurrence of rtz in a micrographic mesostasis, as already mentioned. The plagioe is lath-shaped and frequently intergrown with ophitic augite. hough this quartz-gabbro is part of the so-called hypersthene-gabbro rusion, the marginal facies are very deficient in the hypersthene. ce the petrology of the gabbro is going to be described fully in a later ount, no further details will be given here.

'lear evidence bearing on the true nature of the granophyre which ns the cores of the xenoliths is afforded by a curious 'breccia' lying w yards outside the gabbro boundary at Rudha Carrach (loc. 2, fig. 1). s rock consists of very pale angular fragments set in a uniform, finely stalline matrix. The fragments measure from a fraction of an inch wo inches in diameter. They are composed of recrystallized sutured rtz grains, which are penetrated by irregular veinlets of cryptostalline felspar and quartz. The quartzo-felspathic patches swell out places without disrupting the surrounding quartz aggregate, and it is llent from this and from the increase in the amount of the felspathized terial towards the margins of the angular fragments, that this is a e of metasomatism of a highly siliceous rock. The original rock was ndstone with, probably, a small amount of calcareous cement, as gested by the calcite in the norm of the granophyre (table I) and by occasional patches of calcite which occur in the xenoliths. There is a very small amount of prehnite present in the latter.

n intimate and rather irregular intergrowth of quartz and felspar urs at the margins of some of the sedimentary quartz grains, and asionally the latter are in optical continuity with the lace-like elopment of quartz in the surrounding intergrowths. Generally, wever, the quartz intergrown with the felspar occurs as minute lath-

shaped crystals of random orientation set in a matrix of cryptocrystalline turbid felspar. These may well be pseudomorphous after tridymite, and in that case they indicate that the felspathization occurred at a high temperature, probably well above 800° C. The general appearance of the rock is very similar to that of felspathic Torridonian sandstone which has been metamorphosed in contact with an ultrabasic intrusion on the island of Rhum, and which Harker describes as reproducing 'all the features of granophyre and quartz-porphyry', except in regard to

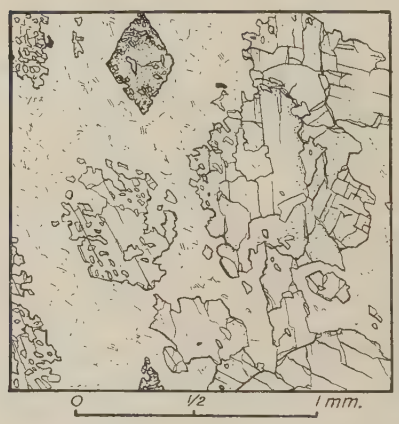

Fig. 2

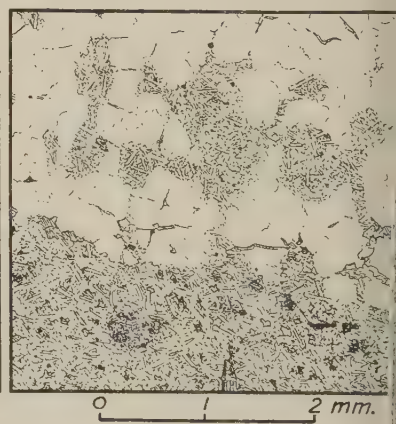

Fig. 3

Fig. 2. Marginal zone of granophyric xenolith, Rudha Carrach (locality 1), showing poikiloblastic crystals of sphene and zoned green pyroxene in a fine-grained groundmass of plagioclase.

Fig. 3. Felspathized sandstone, Rudha Carrach (locality 2).

the preponderance of quartz over felspar (4, pp. 68–69, and fig. 21A). The Rhum and Ardnamurchan rocks differ in one important respect: the felspar in the altered Torridonian is believed to be entirely original, while in the Rudha Carrack rock it is apparently all derived with the aid of metasomatism. The texture produced initially by the felspathization is not truly granophyric for the irregular quartz-felspar intergrowths are not geometrically perfect, as they are in the true granophyre found immediately adjacent to the hypersthene-gabbro and in the xenolith. Small felspar crystals showing albite twinning occur together with the turbid (presumably alkali) felspar intergrown with the quartz in the breccia. The only other minerals present in this rock are minute scales of dark green chlorite, some iron ore, and needles of apatite.

Mobilization of the Rudha Carrach 'granophyre' is shown by the reaking-up and movement of the angular quartz fragments in the reccia. A chloritic streakiness wraps around some of these fragments ke flow lines around the phenocrysts in a lava. Such mobilization ems to be a normal sequence to felspathization: it is shown by grano-yre veinlets emanating from siliceous xenoliths which have been lspathized in contact with lamprophyres in dikes in the Ards peninsula Northern Ireland, and by syenitic veins occurring in hornblendite the coast of Colonsay, Scotland, both described by D. L. Reynolds pp. 367-407; 10, pp. 51-76); while Walker and Poldervaart (14, 614) have described several cases where sedimentary inclusions in the rroo dolerites have been felspathized and rendered intrusive. The e with which this mobility is achieved may be due to the attainment a near-eutectic composition as a result of the metasomatism. This uld account not only for the fluidity, but also for the ultimate pro-ction of a 'micrographic' texture, for it is possible that the latter can ly be produced with any degree of perfection at the threshold of lting of the quartz-felspar mixture. The chemical aspects of fels-thization of highly siliceous rocks have been fully described by D. L. ynolds (loc. cit.). The rock sample from Rudha Carrach selected analysis (table I, A) contains some of the unaltered quartz of the dstone, so that direct comparison with other true granophyres is ficult to make. The analysis of a felspathic rim of a quartz xenolith m one of the basic dikes of the Ards peninsula, formed in somewhat ilar circumstances to the Ardnamurchan rock, is quoted for com-rison.

A diagrammatic summary of the complete sequence of metasomatic cts seen in the xenoliths and at the gabbro contact near Rudha rrach is shown in fig. 4. It is based upon the assumption (which can-be verified) that the original sandstone was homogeneous and com-sed wholly of quartz.

Sedimentary xenoliths in contaminated norite.

A few small pale inclusions of sedimentary origin, with fine laminae hed out by weathering, occur in gabbro which is locally rather pathic and variable in appearance, near the extremity of Sanna nt (loc. 3, fig 1). The xenoliths are composed largely of basic andesine trongly zoned minute crystals with a square cross-section, and xeno-s are surrounded by an ultramafic rim an inch or so wide, rich in

TABLE I. Chemical analysis of felspathized sandstone.

	A.	B.
SiO_2	70·78	70·45
Al_2O_3	11·46	13·14
Fe_2O_3	1·14	1·03
FeO	4·63	0·71
MgO	1·27	1·29
CaO	3·06	2·58
Na_2O	2·21	4·47
K_2O	2·85	3·65
H_2O+	0·98	1·30
H_2O-	0·20	0·34
CO_2	0·39	—
TiO_2	0·92	0·57
P_2O_5	0·15	trace
S	n.d.	0·17
MnO	0·17	0·31
	100·21	100·01

Norm of A.

Quartz		32·10
Orthoclase		17·24
Albite		18·34
Anorthite		13·07
Diopside		1·26
Hypersthene*		
$MgSiO_3$	4·20	10·01
$FeSiO_3$	5·81	
Ilmenite		1·67
Magnetite		1·62
Apatite		1·01
Calcite		0·90

* This represents chlorite in the mode.

A. Felspathized sandstone, Rudha Carrach (loc. 2), Ardnamurchan. Analyst, W. H. Herdsman.

B. Felspathic rim of quartz xenolith in the Kircubbin dike, Ards peninsula. Analyst, E. Kroupa (Reynolds, 1938).

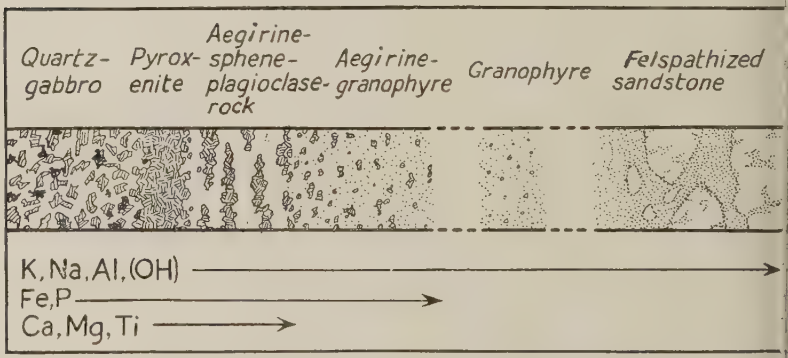

FIG. 4. Composite diagram to show the effects of metasomatism at the contact of gabbro and sediments, Rudha Carrach. The left-hand side of the diagram relates to the xenoliths (see text) and the right-hand side to rocks outside the gabbro contact. The limits of diffusion indicated by the arrows are qualitative only.

hypersthene, followed by two or three feet of norite[1] which grades outwards into the normal gabbro.

In one of the xenoliths, traces of the original bedding are shown by layers of small hypersthene and magnetite crystals, and occasionally by

[1] That is, using the definition of norite as a plagioclase-pyroxene-rock in which hypersthene exceeds clinopyroxene in amount.

enticles of more coarsely crystalline plagioclase (the unstippled layers hown in fig. 5). Occasionally the smaller plagioclase grains are enclosed n poikilitic patches of quartz. The most prominent bedding feature is he ferruginous layer shown in fig. 5. This must have provided an easy ath for the meta-omatizing agents, for he densely concentra-ed magnetite grains re enclosed in large rystals of hyper-hene, mica, and pla-oclase.

Fig. 5. Drawing of polished surface of xenolithic fragment, showing lamination parallel to a prominent layer of recrystallized iron ore, which is invaded but not displaced, by a tongue of mica-hypersthenite. x marks the position of the thin section drawn in fig. 6, Sanna Point.

The ultramafic zone composed of hyper-hene ($En_{65}Fs_{35}$), ark mica, and iron re, with subordinate terstitial patches of uartz having straight oundaries. As shown in fig. 5 a lobe of this mica-hypersthenite threads s way across the bedding planes apparently without disrupting them, nd, when it breaks through the ferruginous layer, the magnetite of e latter remains in place as large irregular grains in the hypersthenite, nd in clusters of enlarged rounded grains concentrated in the cores of agioclase crystals (fig. 6). The latter form part of a very narrow lspathic rim which sheathes the hypersthenite, as shown in fig. 5.

The norite which encloses the xenoliths is a text-book example of this ck-type, as seen from the mode given in table II. Texturally it has ost of the features of a rather fine-grained gabbro, except that the ib-ophitic hypersthene crystals are abnormally prismatic (fig. 7), and e small amounts of mica present frequently surround grains of ore as gged coronas. Only its field association, surrounding the xenoliths, nfirms that this almost ideal orthonorite is a product of contamination the gabbro magma. An analysis of this rock, together with its mode, given in table II. It compares fairly closely with that of a quartz-rite from the Haddo House district, Aberdeenshire.

The only other rock of truly noritic composition so far discovered the hypersthene-gabbro occurs as angular blocks scattered thin a small area on the north-facing slope of Stacan Dubha (loc.). This is a pale, fine-grained felspathic rock, with the hyper-

sthene concentrated into irregular blebs and lenticles, resembling the structure shown in the lower part of the exposure illustrated in fig. 9. Mineralogically this rock has many features which link it with the Sanna Point xenoliths; there is a similar development of minute equant plagioclase grains, with occasional porphyroblasts showing parallel sides but

Fig. 6

Fig. 7

Fig. 6. Part of sedimentary xenolith from locality 3, Sanna Point. The minerals shown are plagioclase (as small grains on the left of the diagram, and as porphyroblasts, bottom right, with magnetite inclusions and ragged terminations); hypersthene (stippled); mica showing cleavage; quartz intergrown with the mica of the top right corner; magnetite both as small grains in the ironstone layer (bottom left) and recrystallized patches in the hypersthenite.

Fig. 7. Quartz-mica-norite from the same locality as fig. 6.

irregular lobate terminations; and quartz occurs in small poikilitic patches in both rocks. The most striking feature of the Stacan Dubha rock is the development of sieve-textured prophyroblasts of hypersthene as shown in fig. 8. All of the hypersthenes occurring in the hornfels and contaminated rocks associated with the hypersthene-gabbro have been found to be more strongly pleochroic, and to have a larger optic axial angle than their gabbroic counterparts. In other words, the hypersthene produced under metamorphic conditions appears to be consistently richer in the ferrosilite component than that produced under magmatic conditions.

The sedimentary origin of the Sanna Point xenoliths with their altered

TABLE II. Chemical analysis of quartz-norite.

	A.	B.
SiO_2	54·02	53·80
Al_2O_3	17·51	15·36
Fe_2O_3	1·14	1·07
FeO	6·30	10·67
MgO	6·73	4·91
CaO	8·42	7·02
Na_2O	3·18	2·50
K_2O	0·19	1·03
H_2O+	1·08	0·60
H_2O-	0·39	0·10
CO	nil	n.d.
TiO_2	0·58	2·60
P_2O_5	0·04	trace
MnO	n.d.	0·15
[illegible]	0·27	0·25
	99·85	100·06

Norm of A.		
Quartz		3·96
Orthoclase		1·11
Albite		27·25
Anorthite		28·30
Diopside:		
$CaSiO_3$ ...	2·20	
$MgSiO_3$ ...	2·00	... 5·65
$FeSiO_3$ ...	1·45	
Hypersthene:		
$MgSiO_3$ ...	14·80	
$FeSiO_3$ ...	8·98	... 23·78
Ilmenite		1·22
Magnetite...		1·62
Apatite		0·34

Mode of A.	
Quartz	7·29
Plagioclase (An_{47}) ...	59·20
Augite	4·62
Hypersthene (En_{65}) ...	25·50
Biotite	1·32
Ore	2·07

A. Quartz-norite, Sanna Point (loc. 3) Ardnamurchan. Analyst, W. H. Herdsman.

B. Quartz-biotite-norite, Quitquox, half a mile north-west of Airdlin Methlick, Aberdeenshire. Analyst, W. H. Herdsman. (H. H. Read, 8, p. 609.)

nstone layers cannot be questioned, and no one studying the structures the Stacan Dubha rock could doubt that it also has a similar origin. e problem of these felspathic noritic rocks is the same as that which nfronted H. H. Read in the case of xenoliths in the Haddo House rite of Aberdeenshire. Read (8, p. 629) concluded that the chances re strongly against the incorporation of country-rock of exactly the ht composition to yield 'micronorite' with hypersthene identical with t found in the enclosing norite, unless some metasomatism had taken ce. If one assumes that the original sediments were pelitic, then the nsfer necessary to bring the mineral phases of the xenoliths into nilibrium with the magma would involve the addition of Ca and Mg, l the reversal of the original K:Na ratio in the country-rock.

Aluminous hornfelses of sedimentary origin.

Hornfelses rich in aluminium and ferric iron are rare in the hypersene-gabbro, but they merit attention since they include the famous undum (sapphire)-hercynite-magnetite-hornfels of Glebe Hill, described in detail by Richey and Thomas (11, pp. 233–235). A second small

group of xenoliths rich in spinel and magnetite, but without visible corundum, has been found by the writer at the base of Stacan Dubha (loc. 5, fig. 1). The chief interest in this new occurrence lies in the structure of the rock, which is well laminated in places, with the ore and

Fig. 8. Hypersthene-plagioclase-hornfels found as scattered blocks on Stacan Dubha (locality 10). The minute plagioclase crystals display the same texture as that found in other xenoliths, e.g. forming the groundmass of the rocks drawn in figs. 2 and 6. Poikilitic patches of quartz are heavily outlined.

spinel laminae projecting, and the intervening plagioclase bands recessed on the weathered surface. Sometimes the structure is more cellular, with a very complete segregation of the dark minerals. This can only be a replacement structure inherited from a well-laminated deposit, and it rules out any suggestion of an igneous origin. Structurally the Gleb Hill hornfels is similar to, though it is more irregular than, the Stacan Dubha example, in respect of the segregation of the light and dark components.

Under the microscope the hercynite in these hornfelses is almost opaque, only the smallest octahedra and the edges of the larger grains showing the deep green colour. The magnetite and hercynite are either densely concentrated to form quite opaque aggregates of irregularly intergrown crystals, or are enclosed in plagioclase which is normally finely granular; but where the opaque minerals are less abundant, the

agioclase occurs as larger crystals, the cores of which are clouded by riads of exceedingly minute, rod-like inclusions of ore. The plagio- se is labradorite, and is therefore not exceptionally aluminous. The owth of the spinel has affected the adjacent plagioclase in places, ing a narrow shell of felspar of different composition from the average, that when the plagioclases are in the position of mean extinction, a rrow bright zone surrounds each spinel crystal. Some of the larger nels act as nuclei for a radial growth of minute felspar crystals. These ks, then, have unusual composition and structure, and from their very normality it should be possible to arrive at a conclusion regarding their gin. The choice seems to lie between two alternatives. In the first ce the concentration of aluminium and ferric iron suggests derivation n a residual weathered deposit, such as a 'bole' as suggested by H. Thomas (11, p. 235). In that case one has to account for the sence of Si in all the minerals except the magnetite and sparse undum, of Ca and Na in the plagioclase, and of Mg in the spinel. an alternative, one may regard these peraluminous hornfelses as the ducts of sedimentary rocks, originally moderately aluminous, to ich Al has been added from the magma. This is the hypothesis oured by H. H. Read (7, p. 453). The Ardnamurchan hornfelses pro- le little fresh evidence towards the solution of this problem. All that e can say with certainty is that the hornfelses have structures of dimentary type, and that, whichever of the above hypotheses is the rrect one, it seems inevitable that some reaction has taken place tween the xenoliths and the magma.

Basic granular hornfelses.

The most abundant inclusions in the hypersthene-gabbro are rocks mposed of the normal gabbroic suite of minerals and having a grano- stic texture. There is at present no satisfactory comprehensive term ich can be used to describe all these rocks. In the past they have en referred to as pyroxene-granulites by Harker (3, p. 115); as gabbro- basalt-, &c.)granoblasts by Schwartz (13, p. 117); and as pyroxene- nulite-hornfelses by A. G. MacGregor (6). Other terms such as erbachite and gabbro-aplite have been used to describe similar rocks ving either a real or deceptive intrusive appearance in the field nelius, 1, p. 4). Objections can be raised to all of these terms on one other of the following grounds:

(*a*) The parent rocks of many of these inclusions are unknown and n only be inferred. This point is clearly brought out by the use of an

analysis of one of these rocks occurring in the Duluth gabbro, in two quite different contexts: firstly, by G. M. Schwartz (13, p. 120) as an indication of the negligible chemical change undergone by basalt in the process of granulitization; and secondly, by F. F. Grout (2, pp. 675–694, anal. no. 13) to illustrate the drastic chemical alteration undergone by slates under the same conditions. Hence the terms gabbro-, or basalt-granoblast are inadmissible except where the origin can be proved beyond doubt.

(*b*) The word 'granulite' without qualification, has been applied to a special kind of quartzo-felspathic tectonite, and its wider use is open to criticism. The present rocks are non-tectonites.

(*c*) Not all of these rocks contain pyroxene: olivine, for example, is dominant in many of them.

The comprehensive term 'basic granular hornfels' is therefore suggested, and for greater precision in any particular case, the leading minerals present may be noted, thus: granular hypersthene-plagioclase-hornfels. The nomenclature of these rocks would then be on the same footing as that of other metamorphic rocks, such as garnet-mica-schist or andalusite-cordierite-hornfels.

In the hypersthene-gabbro these rocks fall into two categories: (1) pyroxene-rich hornfelses with no marked structural characteristics, and (2) layered hornfelses often deficient in clinopyroxene, but containing olivine and hypersthene in abundance. Their distribution is shown in fig. 1, and it should be noted that the layered hornfelses (localities 6–11) are mostly found in the southern part of the gabbro, not far from the contact with Jurassic sediments, while the unlayered inclusions are most abundant towards the west. Only the larger clusters and individual xenoliths are indicated on the map.

The chief characteristics of the two types of inclusion may be summarized as follows:

1. *Granular pyroxene-hornfelses.*—These occur as homogenous masses having a fine, equigranular texture for all components. The only exceptions to this even granularity are provided by relict plagioclase phenocrysts and recrystallized amygdaloidal aggregates, both of which indicate derivation from primary igneous rocks. The granular pyroxene-hornfelses are generally dark rocks containing clinopyroxene in excess of other coloured silicates. There is never more than a feeble segregation of the various components into distinct layers. The surrounding gabbro maintains its normal character right up to the contacts with the hornfel masses, indicating that the incorporation of the latter did little to upse

e balance of phases crystallizing from the magma, and that metaso-atism was probably negligible.

2. *Banded granular hornfelses.* In these rocks both megascopic and icroscopic layering is prominent. The layers may be several inches ick, and in that case they can be traced without discordance to the nits of available exposures which in some places amount to about

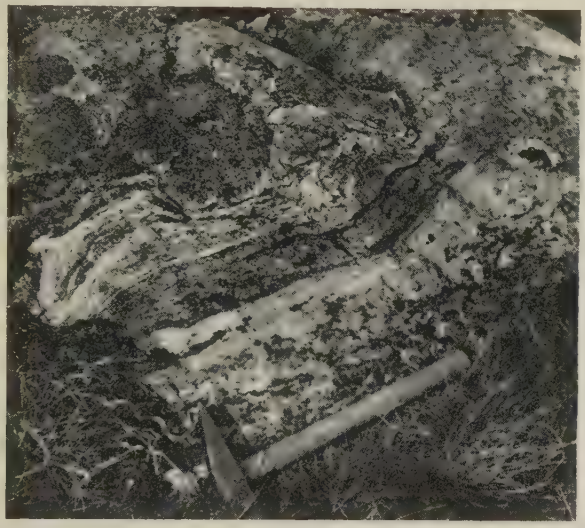

Fig. 9. Typical banded granular hornfels showing segregation of coloured minerals (mainly olivine) into clots. Part of a long narrow strip of hornfels from locality 7, Stacan Dubha.

feet. They are frequently buckled and tend to thicken towards the sts of the miniature folds (fig. 9). Texturally each layer may be stinct, and each component may have its own mode of development: us olivine and hypersthene frequently form sieve-textured porphyro-asts; magnetite occurs as octahedra embedded in plagioclase, or casionally as irregular grains with narrow coronas of pyroxene, amphi-ole, or mica; and the ubiquitous plagioclase is often fluxioned. The abbro' surrounding these inclusions is generally very variable in mposition and there may be narrow transition zones of anorthosite, octolite, leuconorite, &c., between the normal gabbro and the xenolith. ne chief mineralogical feature to note is the subordinate role often ayed by clinopyroxene in these layered hornfelses.

One of the most interesting of the banded granular hornfelses is a

horizontal slab-like mass exposed on a vertical rock face by a fault-gully (loc. 9), a yard or so north of the point where the main footpath south of Beinn na Seilg descends into the gully. It is a dark ferruginous mass about 20 feet long and 3 feet deep, with prominent magnetite layers forming the uppermost 6–9 inches, and with a more homogenous granulite layer below, about 18 inches thick, composed chiefly of plagioclase and magnetite. A photograph of this inclusion is given in A. G. MacGregor's paper[1] (6), showing drag-folding of the upper surface of the inclusion and also variation in the composition of the gabbro adjacent to the inclusion. The normal gabbro consists of labradorite (60 %), augite (20 %), olivine (5 %), hypersthene (4 %), and magnetite (1 %), and towards the inclusion this passes into a zone of augite-troctolite about 6 inches thick, with 17 % of olivine and only 8·5 % of augite. Immediately adjacent to the inclusion proper there is a narrow zone of anorthosite. The following sequence of layers from the top of the inclusion downwards illustrates a very small part of the type of variation that may be seen in many of these rocks (fig. 10):

1. 4 mm. layer of fluxioned plagioclase.
2. 3 mm. layer of similar felspar with tenuous ophitic olivine grains.
3. 1·5 mm. layer of plagioclase with trains of euhedral magnetite grains.
4. 4 mm. layer with large sieve-textured olivine crystals and plagioclase.
5. A layer rich in magnetite rimmed by coronas of hypersthene in some cases and colourless amphibole in others.

No two of these banded rocks are alike. Associated with the hercynite-bearing hornfelses already described (p. 725) there are olivine-plagioclase-hornfelses, with a development of poikiloblastic crystals rivalling those shown in fig. 10. On Stacan Dubha (loc. 7) a slender belt of banded hornfels can be traced intermittently for over a hundred yards, with a characteristic development of the coloured minerals in clots, as shown in fig. 9. West of Beinn na Seilg (loc. 9) there is another occurrence of banded olivine-plagioclase-hornfels with the olivine this time in the form of small, somewhat rounded grains, and augite as occasional poikiloblastic crystals enclosing minute granules of plagioclase and olivine.

Another striking example of banded hornfels is provided by a strip

[1] The choice of this particular photograph to illustrate a paper devoted to pyroxene-granulite-hornfels was perhaps not a happy one, for pyroxene is a very minor constituent of this inclusion.

ich is less than a foot in width, exposed for several yards on the shore rth of Sanna (loc. 11). This has a steep dip, and lies parallel to the ghbouring contact of the great eucrite. The layers are regular and nformable, though slightly puckered in places. They are of two orders

Fig. 10

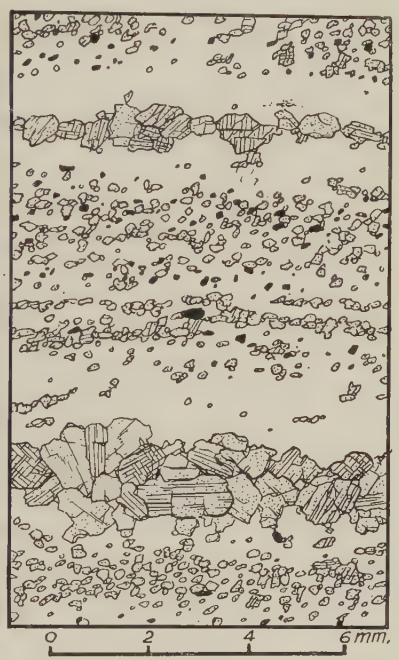

Fig. 11

Fig. 10. Banded granular olivine-hornfels, locality 8, south-west of Beinn na lg. The large sieve-textured patch of olivine is formed of one crystal. The gioclase crystal boundaries have been omitted for the sake of clarity. They are htly fluxioned parallel to the banding, and their size can be judged from the itic intergrowths with olivine shown near the top of the diagram.

Fig. 11. Banded diopside-bytownite-hornfels, locality 11, north of Sanna. The lines of the plagioclase crystals have been omitted. They have an average meter of about 0·5 mm.

magnitude; the finer laminae are no more than a millimetre thick, ile the stouter layers are spaced at intervals of about a centimetre, s spacing being so regular in some cases that it gives a strong sugges- n of a rhythm. The proof that this rock is of sedimentary origin lies in composition, for it is, in effect, a calc-silicate hornfels, with alterna- ns of basic plagioclase (the most calcic to be found in any of the

inclusions, with a composition of about An_{83}) and diopsidic augite. The latter is pale green and non-pleochroic, and its refractive indices (between about 1·68 and 1·70) and 2V (45°-±2°) indicate a composition approximating to $En_{55}Fs_{10}Wo_{35}$. Small and irregular patches of epidote occur in association with the pyroxene, and also along shear-planes. Euhedral crystals of magnetite are fairly evenly scattered throughout the plagioclase-rich layers. The contrast between adjacent layers has been accentuated by diffusion of Fe and Mg to reinforce the pyroxene concentrations, resulting in the removal of all trace of coloured minerals on either side of each prominent dark band through a thickness of about a millimeter. Thus each pyroxene layer is flanked on both sides by narrow bands of pure anorthosite (fig. 11).

The special significance of this particular xenolith is that it displays the same granoblastic texture, the same degree of crystallinity, and the same style of banding as many of the other hornfelses described; but it also has—what the others lack—a composition that is closer, perhaps, to that of certain sedimentary rocks than to any normal igneous ones. Some marls and calcareous mudstones, for instance, have a high lime and alumina, and low silica content comparable with that of the hornfels whose analysis is recorded in table III.

TABLE III. Analysis of banded calc-silicate hornfels, Sanna Point (loc. 11), Ardnamurchan. Analyst, W. H. Herdsman.

SiO_2	...	...	46·26	H_2O+	...	...	0·7
Al_2O_3	...	...	24·18	H_2O-	...	...	0·1
Fe_2O_3	...	...	1·38	CO_2	...	...	nil
FeO	...	...	3·96	TiO_2	...	...	0·1
MgO	...	...	3·58	P_2O_5	...	...	nil
CaO	...	...	17·67	MnO	...	...	trac
Na_2O	...	...	1·54				
K_2O	...	...	0·09				99·7

Origin of layered granular hornfelses.

Four possibilities may be considered concerning the origin of these layered hornfelses.

(*a*) Shearing of the gabbro accompanied by complete recrystallization giving a kind of 'tectonic unmixing' of the various components, such as one may find between the micaceous and the quartz-felspar components in some schists. This mechanism is unlikely to be effective with the gabbroic mineral suite, and even if it were capable of producing a kind of layering, it would not be expected to produce the textural variation shown by adjacent bands in these hornfelses (cf. figs. 5, 9, 10, and 11

addition, many of the banded rocks show none of the mineral grain ntation that one would expect to find if the banding were of hanical origin.

) Thermal metamorphism of banded gabbro. The hypersthene-bro is prominently banded in places, especially near its inner contact n later 'Centre 2' intrusions. Since some of the granular pyroxene-nfelses may be derived from gabbro (see p. 728), it is natural to con-r a possible connexion between banded-hornfels and banded-gabbro. ctual fact, however, the scale of layering in the two rock types is not parable: in the gabbros the order of thickness is from a few centi-res to about a metre, while in the hornfelses the laminae may be a millimetre to a few centimetres thick.

) Thermal metamorphism of bedded rocks (sediments or tuffs) of opriate initial composition. There appear to be no sedimentary s which compare closely with these basic hornfelses in composition. rrelation might perhaps be made between the compositions of the ynite-corundum-hornfelses and those of residual deposits rich in inium and iron; and between the diopside-anorthite-rock and a reous sediment such as a calcareous mudstone or even a greywacke. n in these two most favourable cases there are difficulties. In the instance one has to take into account the calcium and sodium of the ar, and magnesium of the spinellid, none of which could be expected y quantity in a residual deposit. In the second case the original ment would need to be abnormally rich in sodium and almost com-ly deficient in potassium, which would be the reverse of the normal ion with rocks containing a pelitic component. The hypersthene-olivine-hornfelses are even more difficult to explain by non-additive morphism.

) Metamorphism and metasomatism of sediments. This is the pre-d hypothesis: it is the only one that can account for the structure of ayering, for the mineral composition within the layers, and for the ls of crystal habit such as the development of sieve-textured hyroblasts. The latter provide evidence that diffusion processes been in active operation.

Conclusions.

was stated on p. 717 that the country-rock into which the gabbro ma was intruded is mainly sedimentary, yet the majority of the liths appear, at least at first sight, to be of igneous origin. The nition of the true nature of the banded granular hornfelses reduces

the extent of this anomaly, although it does not remove it altogether. The paucity of stoped sedimentary country-rock fragments could conceivably be accounted for in part by the difference in specific gravity that exists between fragments of sediment and those of lava, the former perhaps floating and the latter sinking or remaining suspended in the magma at the present level of erosion. Such a sorting action depends upon the buoyancy of rock fragments in a 'magma' of unknown viscosity and an unknown state of crystallization at any stage in its history, and it is doubtful if xenoliths could move very far from their place of origin except with a magma surge, when they would all travel together irrespective of density differences.

It is far more likely that their initial composition has determined the chances of survival of the different types of country-rock. Basalt fragments no doubt survived recrystallization with little change in composition; but the chemically contrasted sediments probably suffered reaction, leading either to complete assimilation or at least to marked change in their composition. This possibility has hitherto not been considered in relation to the British Tertiary gabbros. For the early history relating to the development of such a concept, one has to examine various works on the inclusions in the Frankenstein gabbro of the Odenwald. Here, dike-like masses of pyroxene-hornfels occur which were thought by C. Chelius (1, p. 4) to cut the gabbro, and he described them as gabbro-aplites or beerbachites. E. Kalkowsky was the first to recognize them as inclusions of material older than the gabbro, and later G. Klemm suggested their derivation from various mixed sediments and tuffs (5). These conclusions were questioned by A. G. MacGregor (6) partly because some of Klemm's 'sedimentary' xenoliths had the textural characteristics found in baked lavas of the British Tertiary province; and partly on account of the difficulties of equating the composition of these basic hornfelses with those of known sediments. This conflict of views may be reconciled, (i) if a broad twofold division is recognized in the rocks concerned, as outlined on p. 728; and (ii) if the full impact of metasomatism is realized.

F. F. Grout (2) appears to be the first geologist to have recognized, rather to have assumed, the large scale operation of metasomatism in converting Virginia slates around the Duluth gabbro into 'pyroxene granulite' xenoliths. Although he quotes a number of analyses, there is too great a gap between those of the contact-altered slates, and of the supposedly further altered granulites, to give a convincing picture of the chemical aspects of basification. Unfortunately, the present accou

oes nothing to close that gap; but it does present critical evidence to now that such basification of various types of sediment is indeed a ality.

This conclusion, which is based largely upon details of structure and mposition of isolated xenoliths, is greatly strengthened by the recent scovery made by M. S. Sadashivaiah (12) of layered basic hornfelses in eir proper field setting, interbedded with cordierite-hornfelses, &c., of e Macduff slate group in contact with the Insch gabbro, Aberdeenshire. ese basic hornfelses duplicate many of the features of the xenoliths the Ardnamurchan hypersthene-gabbro, including the development sieve-textured olivine porphyroblasts.

In the present state of knowledge the writer is not prepared to extend e hypothesis of basification of sediments to include any basic horn-lses other than those which carry an imprint of their past history in eir own structure. Future work will probably show that such an tension is necessary, to cover many of the less spectacular 'pyroxene-anulites' which have no conspicuous structures, and which at present e classed as *incertae sedis*, or are assumed to be of igneous origin.

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Cataclastic pegmatites and calc-silicate skarns near Bunbeg, Co. Donegal.

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Introduction and general field relations.

In this paper is described a series of three small pegmatite dikes and the skarns produced by them in Dalradian calc-silicate hornfelses, and after full discussion certain conclusions are deduced as to the mode of origin of these rocks. This suite has not been described previously and lies to the south of Bunbeg quay and to the east of Loughlin island (fig. 1). The only published accounts of the area contain but passing reference to some of the features here described. In 1891 Nolan (1) merely noted the existence of 'remarkable beds of garnetiferous limestone with tremolite' near the small village of Bunbeg, whilst Zealley (2) in 1909 briefly referred to these hornfelses.

The hornfelses form part of an 'island' of pelites, hornfelses, and epidiorites which in plan is completely encompassed by the Donegal granodiorite, so as to appear to be what many geologists would call a roof pendant. The dikes to be described appear to trend east–west, and are seen in a vertical cliff immediately to the east of the southern point of Loughlin island (fig. 1). Both the strike of the metamorphic rocks and the lineation of the granodiorite (as determined by the preferred orientation of the felspars) are sensibly north–south (3). Unfortunately it is not possible to correlate these hornfelses with any others in the area. The limestone of Lough Anure described by Pitcher (4) is an analogous 'island' 4 miles SSE. of Bunbeg, and is one of several similar masses which are 'certainly structurally related to one another' (4, p. 126), and occur along a NW. line. The Bunbeg calc-silicate hornfelses cannot be justifiably correlated with these masses, being over $2\frac{1}{4}$ miles to the NE. of this line, and being separated from them by granite. The only other metamorphic rocks in the vicinity are massive quartzites occurring $1\frac{1}{2}$ miles to the east.

Field relations.

The hornfelses.

In a 21-foot high cliff along the western side of the tidal inlet immediately to the east of Loughlin island is an excellent exposure of the varied calc-silicate hornfelses dipping 75° E. of N. at 45°. The bedding is clearly picked out by compositional variation giving distinct bands of grossular and idocrase set in a white to blue-green matrix composed

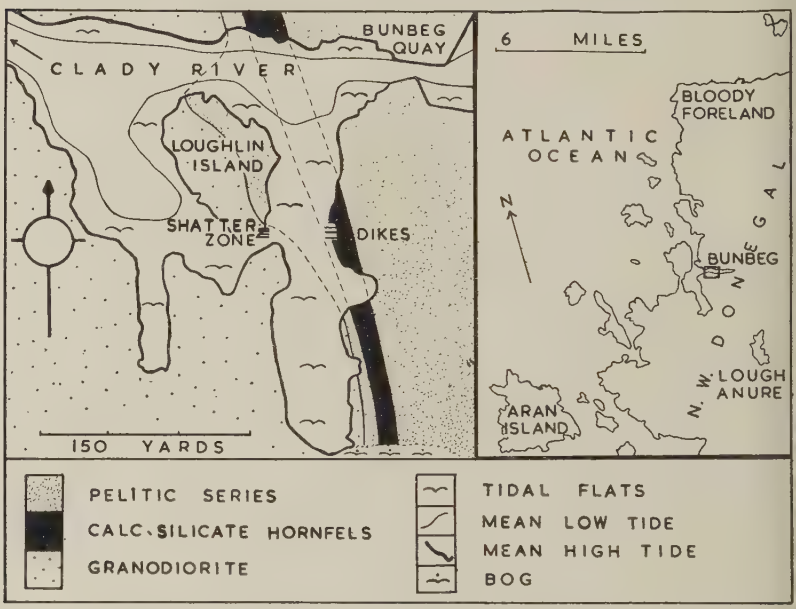

Fig. 1. Sketch-map showing location and geology of the Bunbeg area, Co. Donegal.

primarily of diopsidic pyroxene, wollastonite, and calcite. These minerals can usually be determined macroscopically. The formation, however, is by no means uniform, there being a general tendency to pass from a definite banded rock at the base of the exposure up into a more massive hornfels. Throughout the whole series there are isolated rhombic dodecahedra of garnet intergrown with idocrase, whilst in the lower part of the cliff-section besides these aggregates the bedding planes are picked out by bands, often 2 inches thick, of almost solid garnet and idocrase. In both cases there is an intimate intergrowth of these two minerals, and in the former case the idocrase actually occurs within the crystal boundary of the garnet. The higher, more massive pale greeny-blue

rock only contains the euhedral garnets (or garnet-idocrase intergrowths) scattered randomly through it, whilst wollastonite needles appear lying in the bedding planes at most horizons.

The hornfels has apparently been stretched along a north–south axis, as in the lower part of the exposure the more competent garnet-idocrase 'beds' have developed small boudinage structures, and the adjacent hornfels clearly shows compensational flowage (see fig. 2), and other internal plastic flow-structures occur at the same horizon.

Fig. 2. Boudinage structure developed in a competent garnet-idocrase band in the calc-silicate hornfels. The less competent bands, rich in calcite, diopside, and wollastonite, are seen to have effected compensational flowage. $\times\frac{1}{2}$.

The pegmatites.

There are three principal dike-like masses to be seen in the cliff, but observation is limited to two dimensions, which renders interpretation difficult. It is thought that these composite pegmatite dikes represent two phases of intrusion (distinguished in the subsequent text and in fig. 3 as I and II) and that their structure has been complicated by a complex cooling history which induced a zonal arrangement in the rocks of the second intrusive phase (II). Thus in the largest and most northerly dike, $4\frac{1}{4}$ feet wide at the base of the outcrop and tapering upwards a few inches, the sequence shown in table I can be detected. The spatial relations of the rock-types are indicated in fig. 3. It can be seen that the second intrusion was emplaced on either side of the initial pegmatite. The zonal arrangement is most clearly seen on the south flank of the dike, whilst on the north flank general decomposition, rotting, and crushing obscure the contacts.

This time-sequence between the two intrusive phases (I and II) is maintained because irregular stringers penetrate into the central material of the first intrusion (I), and also a xenolith of the latter is incorporated within the coarser medium-grained pegmatite of the second

intrusive phase (II*c*). These relations are schematically illustrated in fig. 3B.

The second intrusion (II), where it is in contact with the hornfelses, has developed dark reaction zones varying in width from about $\frac{1}{4}$ inch to nearly 2 inches. The hornfelses have been differentially transformed,

TABLE I. The order of intrusion and consolidation illustrated in the largest pegmatitic dike.

Intrusion.	Description.	Grain-size.*
I	Diopside-microcline-perthite-pegmatite	Fine
II	Crystallized to give successive zones as follows:	
II*a*	Contact skarn zone	Fine
II*b*	White felspathic zone	Fine
II*c*	Medium pegmatite zone	Medium
II*d*	Quartz zone or core	Medium

* Fine, < 1 inch; medium, 1–4 inches.

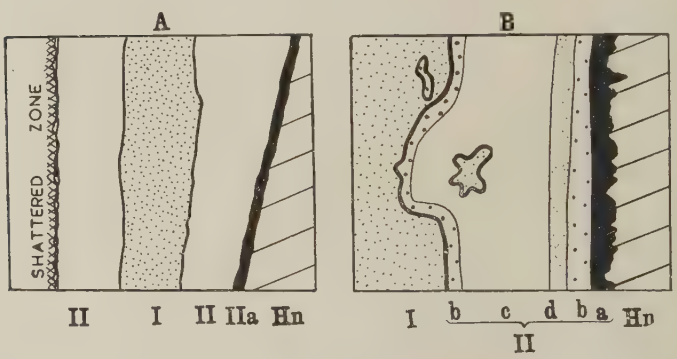

FIG. 3.

A. Diagrammatic cross-section of the largest dike.
B. Detailed section of the right-hand side of the same dike

Hn, Hornfels. I, First intrusion. II, second intrusion. II*a*, Contact skarn zone II*b*, White felspathic zone. II*c*, Medium pegmatite zone. II*d*, Quartz zone.

as some of the more compact garnet-idocrase bands form small promontories projecting into the reaction skarn (II*a*). There is a sharp boundary between this reaction material and the cataclastic white felspathic zone (II*b*), but the latter blends into the central fine pegmatite of the first intrusion (I). It is postulated that, as shown in fig. 3B, the exposure shows a section across a sequence of successively precipitated zones of material deposited in a rudely concentric manner upon the then

confining walls. Hence the material designated (II*b*) is considered to represent the first consolidation from the second magmatic intrusion. The medium pegmatite (II*c*) then consolidated, and the last fraction to solidify comprises the quartz zone (II*d*) representing the 'core' of the intrusion. As at present exposed the penultimate zone to consolidate, which is about 9 inches wide and contains felspars up to 3 inches in length set in quartz, is seen to be incomplete because the quartzose core (II*d*), which is $1\frac{1}{2}$–6 inches wide, has consolidated adjacent to the white felspathic pegmatite (II*b*).

The initial fine pegmatite (I) is of variable width (reaching $2\frac{1}{2}$ feet), and on the north side passes into medium textured pegmatite of the second (II) intrusion. Some 12–18 inches of the latter material is exposed, but as already stated the northerly extension is obscured.

This description applies to the largest of the three dikes, which is the only one reaching the base of the exposure. The other two are situated a few yards to the south, and both of these completely taper away downwards. They possess prominent reaction zones marginal to the milky white intrusive which simulates the white felspathic zone (II*b*) of the largest dike. Still farther to the south there is also a quartz vein, which, though sinuous, bears a dike-like relation to the hornfelses, and appears to have been to some extent differentially emplaced, etching out some of the bedding planes more than others. It is also sheathed in a minute reaction border, which varies 2–7 mm. in width and consists of wollastonite needles arranged parallel to the dike, except in the wider portions of the skarn where they assume a rudely radial orientation. Diopside crystals occur in interrupted thin bands along the skarn. The thin bands of diopside and clinozoisite of the *S*-planes (representing the original bedding planes) die away in the wollastonite-skarn, and it is probable that much of the pyroxenic material of the skarn has been produced from this primary material.

Microscopic characters.

The calc-silicate hornfelses.

As previously noted these rocks show a great deal of variation and are composed of varying amounts of dull red grossular, zoned brown idocrase, pale green diopside, wollastonite, calcite, quartz, and andesine. The proportion of calcite is the most variable; sometimes rocks with a gneissose appearance are composed dominantly of granular calcite (about 1 mm. in diameter) traversed by fine bands of diopside, intergrown

grossular and idocrase, and occasional wollastonite (fig. 4). There is every variation in the amount of calcite present, so that these carbonate-rich bands contrast with those which are essentially grossular-

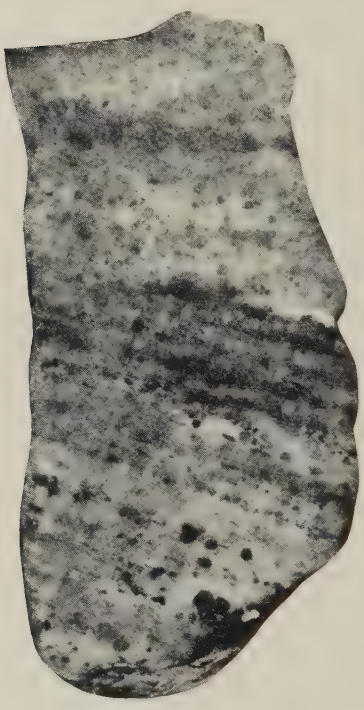

Fig. 4

Fig. 5

Fig. 4. Hornfels composed of calcite with small diopsides dispersed through it, and with the original bedding planes picked out by thin bands of idocrase (black) and grossular garnet (deep grey) which weather into prominent ridges on exposed surfaces. ×1·1.

Fig. 5. Hornfels showing abundant wollastonite (white) associated with quartz and diopside (pale grey), and porphyroblasts of aggregated garnet (deep grey) and idocrase (black). The subhedral porphyroblast is of the type referred to on page 744. ×1·3.

idocrase-rocks. These minerals are usually interspersed with diopside, andesine, and quartz. All the silicate minerals are more abundant along certain *S*-planes, whilst the idocrase is frequently elongated parallel to the *c*-axis, which tends to lie randomly in this plane. Likewise wollastonite is always acicular, with the needles in the *S*-planes. It varies from occurring as scattered needles to dense felted masses as shown in fig. 5

It is thought that the quartz and andesine (Ab_6An_4 by Michel-Lévy method) have resulted from the introduction of exotic materials along the *S*-planes during the emplacement of the regional granite. Both minerals only rank as accessory constituents. Since this quartz is intimately associated with calcite, whilst wollastonite-bearing calcareous bands

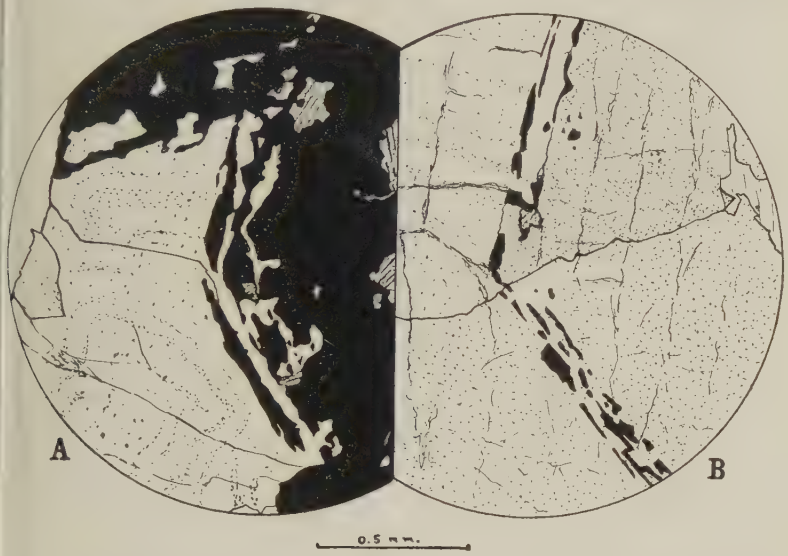

Fig. 6.

A. Hornfels showing differential replacement of garnet zones by zoned idocrase, representing an early stage in production of palimpsest structures. Diopside and quartz present in the garnet.

B. Palimpsest of euhedral grossular from hornfels which has been almost completely pseudomorphed by idocrase.

Crossed nicols. Garnet isotropic, idocrase stippled. ×20.

were produced in adjacent horizons by the initial metamorphism, subsequent metasomatism is indicated. This metasomatism has also effected diaphthoretic breakdown of grossular to idocrase in a manner similar to that described by Osborne in Carlingford (5). Contrary to the condition which so often exists, however, there is in these rocks no doubt whatsoever as to which mineral is being replaced. Thus, for example, within the limits of a single slide two excellent palimpsest euhedral grossulars were observed in differing states of replacement, as illustrated in fig. 6. The idocrase has obviously differentially picked out the zoning of the garnet. Such secondary idocrase frequently has better crystalline shape

than the grossular, and often contains poikiloblastic clusters of garnet, giving a sieved appearance to thin sections. Further, the idocrase has grown about many of the diopsides contained in the *S*-planes without disturbing their positions, which is indicative of differential metasomatic growth of the secondary idocrase.

The grossular is sometimes massive, but more frequently occurs as subhedral to euhedral rhombic dodecahedra up to 1 inch in diameter (fig. 5). Mutual sieving commonly occurs between idocrase and the garnet, and in turn both are often highly sieved with diopside and a few grains of calcite. The poikiloblastic diopside occurs as rounded units normally with random distribution within the garnet, but occasionally a distinct linear band of similar grains may traverse an euhedral porphyroblast without having been deflected from the original *S*-planes by the growth of the grossular. This fact is considered to be of importance in connexion with the method of growth of porphyroblastic garnet in hornfels, as this example clearly shows that the developing crystal although growing in a solid medium, did not express any facility to remove the 'unwanted' materials, neither was the enclosing hornfels deformed. Hence growth in the solid has not, in this case, necessitated the exercise of mechanical disruptive forces during crystallization. The previously described idocrase exhibits analogous relations. It is interesting in contrast to note that Osborne records 'certain garnet-rich rocks where the smaller wollastonite units have been pushed aside by the growth of the grossular crystals. . . . In such rocks the structure is porphyroblastic' (6, p. 294).

The pegmatites: the largest dike.

The mode of a representative sample of the first intrusion (I) is detailed in table II (column 1). The frequently twinned euhedral diopside is a salitic mineral of the diopside-hedenbergite range, occurring as crystals with a maximum dimension of 3·3 mm. Uralitic marginal alteration and patchy replacement by calcite are frequent. The diopside is set in quadrille-twinned microcline and microcline vein- and patch-perthite areas reaching 6·5 mm. in diameter. Some of the quartz was a deuteric introduction, being brought in along inter-crystal boundaries, and frequently producing slightly bulbous replacements, at times resembling crypto-myrmekite. This late-stage quartz entering along fractures in the pyroxene has 'stoped' and floated fragments apart. The quartz and sometimes the felspar, is riddled with strongly prismatic euhedral apatites.

Sphene, which is tinged with pink in thin section, is abundant and perfectly euhedral, in contrast with the anhedral fragments occurring in the skarn (II*a*). This rock would appear to be of the type styled 'sphene rock' by Scott (7) in SW. Donegal, a term endorsed by Scott, Griffith, and Haughton (8), and later referred to by Zealley (2). This name is completely inappropriate in view of the fact that despite the noteworthy abundance of sphene which is a feature of this rock, modally it comprises less than 2 %. The diopside-bearing pegmatites from near Ellon described by Read (9) appear to be very analogous rocks.

TABLE II. Modes calculated from micrometric measurements and expressed as volume %.

Minerals.	1.	2.	3.
Microcline and microcline-perthite ...	57·2	71·7	23·7
Plagioclase	2·7	3·2	65·2
Quartz	26·4	20·8	6·7
Diopside	12·6	4·3	—
Sphene	1·2	present	0·1
Clinozoisite	—	—	2·1
Calcite	—	—	0·2
Apatite	present	trace	—
Muscovite	trace	—	—
Biotite	trace	—	—
'Pyribole aggregate' (see page 747) ...	—	—	1·4
Colour index	13·8	4·5	3·8

1. Diopside microcline-perthite-pegmatite of first intrusion (I) into the largest dike.
2. Pegmatite of second intrusion (II*c*) from north flank of the largest dike.
3. Central pegmatitic material from the middle dike macroscopically similar to (II*b*) of the largest dike.

In the white felspathic zone of the second intrusive phase (II*b*) microcline vein-perthite is the dominant mineral, together with a considerable amount of quartz and diopside associated with some oligoclase and a little sphene (as large euhedral and anhedral grains). Swarms of minute apatite prisms are also present, occurring most abundantly in the quartz. A little deuteric albite occurs around some of the perthites, and is sometimes in optical continuity with the exsolved albite. The diopside is fresh. The quartz is highly strained, although the rock as a whole lacks evidence of cataclasis.

On the north flank of the largest dike occurs a somewhat fine-grained facies of the medium pegmatite zone (II*c*) whose mode is given in table II (column 2). The pink colour of the rock is given by the microcline film-perthite occurring as grains up to 2·6 -5·0 mm. in diameter. The rock shows excellent cataclastic structures with disruption and strain of the

felspars, and there are frequently very ragged boundaries between th quartz and felspar. Small gaps left between these larger crystals o intrusion have been filled by minute quartzose and felspathic blebs Creasing and twisting of quartz grains is common. In fig. 7 a larg sheared sphene is seen adjacent to a highly twisted and mechanicall

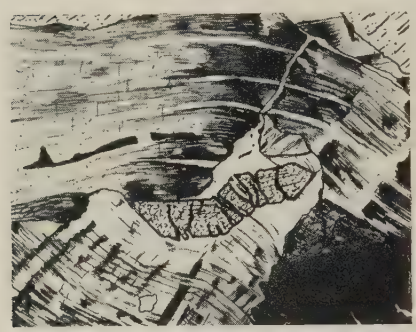

Fig. 7

Fig. 8

Fig. 7. Creasing and twisting of quartz (above) and microcline (below), and di ruption of a sphene (centre) as a result of cataclasis in the pink pegmatite (IIc) the second intrusion into the largest dike. Crossed nicols. ×13.

Fig. 8. Photomicrograph showing unusual spindle-shaped strain-shadows quartz produced during cataclasis of the pink pegmatite (IIc) of the second intrusi into the largest dike. Crossed nicols. ×30.

disrupted quartz. Cracks traversing both the quartz and adjace felspar are filled with calcite. This stage of deformation was after som but before other, quartz crystallization, because similar gashes are som times filled with quartz. Fig. 8 shows the effect of an unusual type quartz straining. This has produced parallel spindle-shaped strai shadows giving irregular partial extinction, which can only be likened the appearance of the twinning characteristic of microcline, but occu ring in only one direction. This, following the results obtained by Grig and Bell (10), would seem to indicate deformation at relatively hi temperatures, moderate pressure, and in the presence of solutio which it is claimed creates a tendency to fracture quartz into needl with definite crystallographic dependencies. Such conditions wou appear to be consistent with those giving rise to the metamorphism the calc-silicate series.

The well-cleaved euhedral sphene crystals are up to 1·5 mm. lor Calcite in addition to forming the veins alluded to above, has almo

completely pseudomorphed the diopsides. The felspars are considerably kaolinized.

The pegmatites: the two smaller dikes.

These two dikes are essentially similar and in both cases the central pegmatitic rock macroscopically simulates the white felspathic zone (II*b*) of the largest dike. A modal analysis of this rock from the middle dike is given in table II (column 3). More than 65 % of a well-zoned basic oligoclase (with excellent albite and pericline twinning) is present, occurring as crystals up to 9 mm. long. This plagioclase shows mortice-structure produced by cataclastic disruption, which gives uneven extinction. The microcline is slightly perthitic and poorly twinned. The uneven grain-size of the rock is due to late packing of the spaces between phenocrysts with smaller (0·1 mm.) grains of plagioclase and quartz. Clinozoisite is quite common, and crystalline calcite is often associated with it. Both these minerals occur interstitially and are bounded by rectilinear margins, taken to indicate late-stage or deuteric development, and since this calcite occasionally cuts through the epidote, the latter was probably solid before the calcite.

The modal 'pyribole aggregate' indicated in the mode is uralitized diopside, with traces of the pyroxene in the core, and in some cases clinozoisite is intimately intergrown with the uralite. Some of these complex masses show alteration to diffuse calcite.

The marginal facies of the middle dike adjacent to the skarn shows diopside-pegmatite with predominant microcline and some plagioclase and clinozoisite. A little replacive, and also some interstitial, calcite occurs together with a trace of fibrous muscovite (the latter having a rather 'zeolitic' habit). The diopside is uralitized marginally. Plagioclase is highly kaolinized, whilst the potash-felspar is fresh and sometimes has untwinned cores. Many of the large felspars have been mechanically disrupted giving anomalous extinction patterns probably due to cataclasis during magmatic intrusion with concomitant 'cementation' by a host of smaller crystals of quartz and felspar. The complex texture makes modal analysis difficult, but potash-felspar predominates considerably over plagioclase.

Contact skarns of the two smaller dikes.

The second pegmatitic intrusion (II) has produced a marked skarn at its contacts with the hornfels. The middle dike tapers downwards to such an extent that near its lowest extremity the intrusion becomes less

than an inch wide, and the pegmatitic material therein comprises about 70 % felspar, there being nearly equal quantities of albite-oligoclase (about Ab_8An_2) and patch-perthite. All this felspar is shattered, with chlorite and calcite filling the ruptures. Quartz, after the felspars, is the next most abundant mineral and occurs as small grains. Both the 5–10 % of calcite and the small proportion of clinozoisite are thought to be primary in view of their straight line contacts and interstitial relations. A little zoned chalcedony exhibits analogous relations.

Cataclastic structure is not so marked in this rock, but mortice-structure in some of the felspars, and disparity in crystal size probably imply late cementation of the larger grains (which experienced slight cataclasis during intrusion of the magma) by smaller second generation crystals with grain-sizes ranging 0·30–0·07 mm. The latter are composed of quartz, lamellar-twinned plagioclase, some untwinned felspar (? K-felspar), together with calcite, clinozoisite, a little rather mushy muscovite, and occasional chalcedony.

In the contact zone there is copious very pale watery-green non-pleochroic diopside showing uralitized margins. Quartz is a little more abundant than in the pegmatite. Felspars (abundantly rich in apatite needles) are, being the finer in grain-size, packed between the pyroxenes. Calcite is quite prominent and occurs both as interstitial masses and also as regular-edged replacements in other minerals; frequently the felspars are almost completely replaced by calcite. Thus calcite has two habits; that which is interstitial, filling angular gaps between minerals, would appear to be primary, whilst the sprawling replacements of felspar, which are more prominent in the contact zone, appear to be of a metasomatic origin.

Contact skarn zone (IIa) of the largest dike.

This skarn may be divided into two distinct reaction zones:

(*a*) the 'black' zone adjacent to the white felspathic zone (II*b*), and

(*b*) the 'green' zone adjacent to the hornfels (Hn of fig. 3).

The white felspathic zone (II*b*) close to the 'black' zone contains plagioclase, quartz, and some microcline, all being highly cataclastic, whilst calcite, clinozoisite, sphene, and muscovite also occur. The plagioclase has a composition between oligoclase and andesine and occurs as grains up to 1·5–2·0 mm. diameter. Anhedral quartzes are sometimes more than 3 mm. long. The clinozoisite and calcite are mainly interstitial and are thus presumed to be primary, although part of the calcite and a

little of the epidote has a nebulous replacive secondary aspect. Some muscovite is also interstitial.

The plagioclase shows marked cataclasis with prominent micro-shears and also distortion by cracking with opening of the cracks allowing entry of clinozoisite and calcite. It is considered that the latter two minerals are primary, as in the case of the Fen carbonatites described by Bowen (11 and 12). It is also noteworthy that according to Eskola (13) the Becket granite-gneiss of Massachusetts can contain 2 % of primary calcite, whilst von Eckermann in his latest work on the Alnö complex accepts the agency of magmatic carbonate (14, p. 99). Sometimes even the calcite appears to be slightly strained giving wavy extinction and twisted lamellae, and likewise some of the interstitial clinozoisite and muscovite is twisted to a limited extent, but none of these minerals has suffered disruption comparable with that manifest by the felspathic minerals.

In the 'black' zone of the skarn biotite and actinolite occur. The actinolite appears as fibrous masses $< 0{\cdot}6$ mm. long arranged in irregular clots along the contact. It has an absorption of $\gamma > \alpha$ (being pleochroic between bluish sap-green and very pale yellowish sap-green), and $\gamma : c$ about 16°. It shows an uralitic habit, sometimes sending fine fibrous masses into the surrounding minerals, and high-power examination shows small-scale alteration to calcite. The biotite occurs in two habits —either as small flakes, or alternatively as sub-radial interstitial material of 'zeolitic' habit, such as is frequently found in metasomatized aureoles. Along this contact, but just within the pegmatite, a few small ragged penninized grains of biotite are developed between disrupted felspars, and are often intimately associated with deuteric calcite as shown in fig. 9. The calcite illustrated is both interstitial and also somewhat replacive with respect to the felspars. The biotites in the skarn show effects of minor mechanical strain which occurred prior to the latest deuteric stage. Fig. 10A shows that mica has replaced anhedral cataclasized oligoclase (which is also calcitized), whilst its opened cleavages have been filled with quartz and possibly a little felspar. These features are reminiscent of the mica-lamprophyre of South Hill, St. Helier, Jersey (described by Searle, 15), in which magnesian carbonate has filled opened mica cleavages.

Sphene is abundant as minute broken or anhedral grains, frequently occurring in swarms. Subhedral diopside occurs sparingly adjacent to the pegmatite, and steadily increases in concentration towards the 'green' zone. Some quartz also occurs in this zone.

Towards the centre of the 'black' zone corroded fragments or palimpsests of silicates, particularly oligoclase occur within areas which have been completely calcitized, as illustrated in fig. 10B. Towards the 'green' zone small apatites become more abundant, and large sinuous quartz crystals also become common, being most abundant along the junction between the zones. The 'green' zone is narrower than the 'black', and is clearly transitional between the latter and the hornfelses. Biotite and sphene, common in the 'black' zone, now disappear, whilst diopside, quartz, and calcite become dominant, and anhedral isolated grossulars increase towards the hornfels. Diopsides of all sizes sieve the calcite, whilst the quartz was probably a late introduction. The resistant bands of hornfels which embay the skarn zone are composed of the usual idocrase-grossular assemblage with a little diopside and calcite.

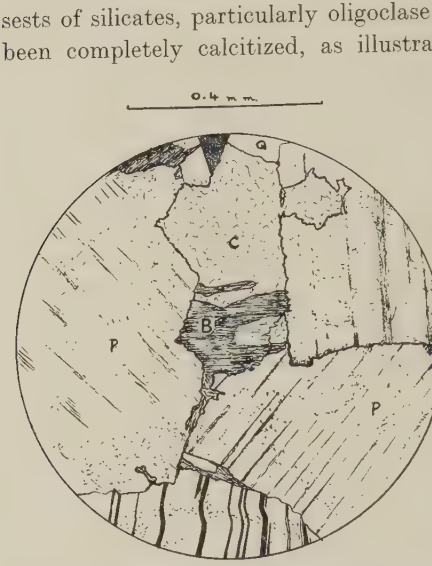

FIG. 9. Pegmatite (II*b*) marginal to the skarn of the largest dike, showing the intimate relationship between biotite (B) and deuteric calcite (C), and also the interstitial and replacive habits of calcite with respect to plagioclase (P). Quartz (Q) is also present. Crossed nicols. ×32.

Although these are the general features of this contact skarn, it should be recorded that near the top of the cliff the dike is considerably broader and consequently the effects of mechanical strain were less intense, and there is a slightly different assemblage. The effects of cataclasis on the felspars seems to be partially crystallographically healed and late-stage quartz has entered between the crystal boundaries, partially replacing the felspars. In the 'black' and 'green' zones relations are similar except that in the former actinolite and biotite are lacking, and in the latter quartz and felspar are more prominent so that a distinct light-coloured rim occurs round the garnet-idocrase promontories. Also the felspar in the skarn has been decomposed, but not so strongly calcitized as in the lower portions of the dike where cataclasis is more marked. Hence it appears that in this upper part late-stage (deuteric) action was curtailed

restricting the formation of actinolite, biotite, and calcite, probably as a result of less easy access consequent upon poorer cataclasis. The introduction of quartz, however, has been intensified, both in the extreme marginal part of the dike and also in the outer part of the skarn.

Fig. 10.

A. The association of plagioclase (P), calcite (C), biotite (B), actinolite and pennine (A), and sphene (S) in the contact skarn II*a* ('black' zone) of the largest dike. The biotite reflects minor mechanical strain which opened the cleavages allowing infilling with quartz.

B. Advanced calcitization of the 'black' zone skarn of the largest dike, showing active pseudomorphism of all the silicates. Note especially the marked replacement of the euhedral plagioclase in the centre of the field. Euhedral diopside (D) abundant. Plane polarized light. ×17.

Intrusion sequence.

The production of skarn is thought to have been a composite process because of the dual intrusion. The initial intrusion assimilated hornfelsic materials, and this augmented magma produced the potassic pegmatite with nearly 13 % modal diopside (table II, column 1). The hornfels may have been completely dissolved prior to the pegmatite crystallization, or alternatively the diopside may simply have been recrystallized. Even the latter hypothesis implies a complete dissolving away of the enclosing hornfels, because the diopside always simulates a

primary crystallization, a feature which is completely analogous to the diopside pegmatites near Ellon, where Read (9) attributes the crystalline perfection of the sphene and diopside to early crystallization from the contaminated pegmatite magma. Campbell Smith (16, p. 181) has suggested that the diopside-bearing veins at 'Marble Outcrop' and 'Marble Dike' in South Victoria Land are similar to these Donegal pegmatites, and in these Antarctic specimens which he described, it is suggested that very acid intrusions assimilated metamorphic limestones and granulites. Emmons and Calkins (17) also record diopside aplites with 7·75 % and 12·67 % normative diopside in respectively scapolite-poor and scapolite-rich dikes cutting the Madison limestone near Philipsburg, Montana. Also in the recent work of Hieke (18) and Schiavinato (19) strong support is given for endomorphic assimilation of limestone by acid plutonic masses (assisted by pneumatolytic-hydrothermal action), giving rise to pyroxene- and sphene-rich skarns.

It is interesting to note that Eskola (13) thinks that 'assimilation usually occurs where the intrusive rock shows banded structure and a high degree of protoclastic crushing and shearing, whilst it does not occur in masses which have been intruded and crystallized under quieter conditions . . .' In describing the Bazena region Schiavinato (19) has also noted the effects of dynamic pressure producing cataclasis in quartz, plagioclase, and calcite of the endomorphic zone.

The diopside-rich first intrusion (I) is only seen in the largest, most northerly dike. The rock-types designated and described as of the second intrusion (II) are all thought to have been part of a later intrusion into the two tapering dikes to the south and also on either flank of the first intrusion (I) in the largest dike. That is to imply that the various rock-types of the second intrusion all crystallized from a single magma and that successive shells or zones were precipitated on the perimeter of the magma cavity. This marked zonal arrangement appears to be in harmony with the recent hypothesis of Cameron, Jahns, McNaire, and Page (20) in which the authors suggest that zoned pegmatites form as a result of fractional crystallization of a single magma intrusion in such a way that successive crops of crystals are deposited against the wall until the chamber is filled. This hypothesis allows for such successive shells being complete or incomplete, due to lack of precipitation or reaction between crystals and rest-liquid. Reference to fig. 3B shows that in the Bunbeg dikes the medium pegmatite zone (II*c*) is incomplete causing the core or quartz zone (II*d*) to be eccentric.

The fact that the two smaller dikes are narrower and tapering would

on this hypothesis, anticipate the fact that the central zones of the second intrusive phase are absent, whereas if a three-dimensional study were possible a more complete series might (and should) be revealed. There are, however, no outcrops immediately to the east of these exposures.

A point of difference exists between the numerous examples described by Cameron, Jahns, McNaire, and Page in their monograph (20) and the largest dike at Bunbeg in so far as in the latter there are two distinct phases of intrusion, whereas in the monograph it is implied that a single intrusion is usual in all pegmatites. A difficulty also arises in connexion with the hydrothermal quartz-vein which occurs farther to the south. The authors of the monograph in enunciating their theory give numerous examples of 'fracture fillings' which are described as being primarily quartzose, and cutting both the parent pegmatites and the country-rocks. It is here suggested that the quartz vein at Bunbeg is such a fracture filling, but the fact cannot be proved, though it seems reasonable to regard all four of these minor intrusions (the three pegmatitic dikes and the quartz vein) as penecontemporaneous and related.

Acceptance of this hypothesis implies that the skarns here described were produced during the second intrusive episode (II), and in this case there is ample evidence, in the cataclasis, of a forceful injection of a true magma (i.e. a mixture of liquid and solid phases), rather than for the existence of a 'pegmatitic thoroughfare' of the type postulated by Quirke and Kremers (21). The effects of cataclasis would not necessarily be preserved in the inner zones of the pegmatite, where slow cooling would allow 'healing' by recrystallization.

The original cavities into which the pegmatites were intruded were east–west minor shatter zones marginal to the Bunbeg 'island', and the continuation of such shatters is seen in, and has probably given rise to, the topographic gap between Loughlin island and the adjacent shore. It is noteworthy that at this point 3 % of modal muscovite occurs in the granite, whilst in neighbouring localities this mineral is too insignificant to figure in the mode. This granite also contains 45 % of quartz and far less plagioclase than adjoining granites.

Thus, although it must be admitted that some of the interpretations are possibly analogous to those styled 'anthropomorphic' by Perrin and Roubault (22), it is considered that these dikes have been produced essentially by forceful intrusion and concomitant deuteric and/or metasomatic action. This contention is maintained despite the fact that the normative felspar of the pegmatites whose modes are given in table II would undoubtedly, when plotted on an Or: Ab: An felspar ternary

diagram, lie well astride, and away from, the cotectic line joining the Or:An and Ab:Or eutectic points as defined by Bowen (23). In describing some zoned pegmatites from the Black Hills of South Dakota, Higazy (24) has given strong arguments in favour of their production by reorganization of, and differentiation from, pelitic and semi-pelitic media. This concept is partly based upon the hypothesis that the rocks could not have a magmatic origin because the normative felspar lies well within the orthoclase field of the ternary diagram, that is, on the potash-rich side of Bowen's eutectic. Bearing in mind the nature of the enclosing hornfelses at Bunbeg it would seem reasonable to assert that they could not yield the requisite materials (particularly K and Na) to form the dikes, and also the other materials giving the acidification of the granite on Loughlin island. It is also considered to be highly unlikely that the pegmatites were 'produced' from the pelites stratigraphically higher or lower than the calc-silicate series, and then moved to their present location by rheomorphic processes. Hence the criteria used by Higazy would appear to be inapplicable to the cataclastic pegmatites under discussion, but, whilst claiming an essentially magmatic origin for these dikes, it should be recognized that it is unlikely that there is a unique solution to the problem of the genesis of pegmatites, and this paper has been written with a view to giving a contribution to the study of pegmatites rather than that of trying to enunciate dicta applicable to pegmatites in general.

The skarns.

Following from the argument above, the skarn bordering the latest intrusions has been produced by interaction between the second intrusion (II) and the hornfelses. There can have been no turbulence during the reaction because of the sharpness of the junction (excepting possibly in the widest portion of the largest dike); but a little mechanical deformation seems to have continued at all stages in the history. This, however, was only of a mild degree, as seen in the opening of the micro-cleavages already described. The 'black' zone seems to have been produced by metasomatic introduction of hydrothermal materials giving rise to plagioclase and quartz. Potash-felspar does not occur but biotite is present in this narrow zone, and has probably developed instead of microcline under the influence of the hydrothermal activity; likewise actinolite occurs at the expense of diopside.

With regard to the calcite and clinozoisite, it is suggested that these are definitely either of deuteric or late-stage hydrothermal origin. I

this regard it is noted that Bowen (11 and 12) does not think that carbonates could have crystallized from a magma, but that they occur in the Fen area as a replacement. In this area Bowen discussed differential calcite metasomatism, whereby microperthite is attacked first, and subsequently aegirine followed by hornblende.

This sequence of differential replacement has not been perfectly followed at Bunbeg. In the largest dike the calcitization has attacked the pyroxene but not the felspar to any great extent. In the reaction zone plagioclase has been altered most, and the pyriboles to a lesser extent, thus giving a close approximation to the examples described by Bowen. At the same time the replacement is undoubtedly guided by the deformation planes set up by the cataclasis. It must also be remembered that there is plenty of interstitial calcite and clinozoisite which would appear to be of at least deuteric origin, and although close similarities exist between these replacements and those described by Bowen, some of the calcite is definitely primary. Emerson (25, p. 19) records allanite-pegmatites and sphene-pegmatites replacing isolated tracts of limestone included in the Becket granite-gneiss. Their 'silicates are rich in calcium and suggest the capture by the heated solutions of a portion of the calcium from the limestone which was being dissolved to make place for the new rock. The rare elements suggest the deep-seated source of the hot solutions.' Eskola (13) also accepts the existence of primary calcite as previously mentioned, whilst von Eckermann (14) postulates a definite carbonate magma. A deuteric medium would appear to be necessitated to explain the late forming primary minerals associated with the present pegmatites and skarns. The deuteric origin of the calcite is further substantiated by the fact that some of the quartz was introduced at an even later stage.

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